VOLUME XXIX NUMBER 4

THE

JOURNAL OF GEOLOGY

MAY-JUNE 1921

DIFFUSION IN SILICATE MELTS

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INTRODUCTION

Many well-known features of igneous rocks are clearly the result of diffusion. The growth of crystals is accomplished largely by the diffusion of material to crystalline nuclei. Inclusions often exhibit solution at their borders and some diffusion of their material into the surrounding magma. These phenomena involve diffusion through short distances and there is no theoretical objection to the acceptance of diffusion at the required rate. At the same time it has often been inferred that large quantities of matter have diffused through considerable distances. This has been done for the purpose of explaining various associations of igneous rocks and especially the formation of basic border phases. Since Becker's destructive criticism of this view[1] many geologists, but not all, have been less willing to assign to diffusion any important rôle in the production of such features.

Becker's objection was based on the extreme slowness of diffusion in all cases where its rate had been determined. The principal data then available referred to the diffusion of salts in aqueous solution, and it was on the basis of reasonable deduction

[1] *American Journal of Science* (4) Vol. III (1897), p. 27.

from these data that Becker's objection was raised. There has been as yet no discovery of a general principle connecting the rate of diffusion of matter in solution with other physical constants. Each substance has its coefficient of diffusivity (in a given medium) characteristic of the substance and determinable only by experiment. The coefficient is the constant factor, k, in the equation expressing Fick's law of the rate of diffusion, viz.,

$$\frac{dc}{dt} = k \cdot \frac{d^2c}{dx^2},$$

and is evidently equal to the number of grams which diffuse past 1 square centimeter of any plane in unit time when the concentration gradient normal to the plane is unity. This is a very small quantity for substances investigated. For most salts in aqueous solution it is of the order 3×10^{-6} in cm.^{2} per second at the ordinary temperature and increases rapidly with rise of temperature; for molten metals it is considerably larger and of the order 3×10^{-5}, again increasing with rise of temperature.[1]

In the case of the quite different type of matter, molten silicates, it is perhaps not probable, but nevertheless not inconceivable, that the coefficient of diffusivity should be fairly large at the high temperatures concerned in spite of the usual high viscosity. Only actual measurements can make us at all certain that we have a firm basis of fact for the discussion of diffusion in molten silicates. Practically no measurements have been made. Endell has demonstrated the fact of the interdiffusion of lime and microcline glass.[2] Schulze has measured the rate of migration of silver ion in glass.[3] Indeed, it is perhaps principally from the problems of glass manufacture that we gain our impressions of diffusion in molten silicates. There diffusion is often excessively slow, but it should be noted that glasses belong for the most part among the very viscous silicate mixtures. On account of the lack of experimental data it was considered advisable to undertake some measurements of the

[1] Roberts-Austin found for the coefficient of diffusivity of gold into molten lead $k = 3.47 \times 10^{-5}$ at 492°. *Roy. Soc. London, Phil. Trans.*, Vol. 187A (1896), p. 383. See also Van Orstrand and Dewey, *U.S. Geol. Survey, Prof. Paper 95-G.* 1915.

[2] K. Endell, *Silikat-Zeitschrift*, Vol. I, p. 195.

[3] *Ann. phys.*, Vol. XL (1913), p. 335.

rate of diffusion in fused rock-forming silicates. Those here described are to be considered of a preliminary nature. They are not devised with the purpose of establishing precise values for diffusion coefficients from which general theoretical conclusions might be drawn, though this is recognized as a desirable ultimate goal. They may serve rather to aid the geologist in deciding what he may and what he may not reasonably attribute to diffusion.

THE METHOD OF EXPERIMENT

The method followed was that of permitting the diffusion against gravity of a heavy liquid placed in the lower part of a crucible into a lighter liquid in the upper part. Ostwald has said that to make accurate experiments on diffusion is one of the hardest problems in practical physics on account of the difficulty experienced in eliminating convection currents. It is to be noted that this is especially true of aqueous solutions. Water is a thin liquid with a relatively high coefficient of thermal expansion. The driving force of convection, viz., difference of density, is therefore large and the resistance to it (viscosity) small. In silicates, however, these conditions are reversed, the viscosity being relatively great and the thermal expansion relatively small. While a small difference of temperature may establish convection in aqueous solutions, it is not to be expected to have a comparable effect in silicates. In the case of aqueous solutions the density gradient resulting from the composition gradient may be very small, but in the case of the silicates used it is quite large. If the relative densities of diopside and plagioclase liquids at high temperatures are comparable with those at lower temperatures, it can readily be shown that it would be necessary to have a temperature gradient of 20° per mm. in order to counteract the density gradient due to 1 per cent per mm. change of composition. Gradients of composition of this magnitude exist throughout most of the period of experiment. Such temperature gradients are, however, entirely lacking; not only this, but it is easy to make the moderate temperature gradient that does exist of such a sign that it acts together with the composition gradient instead of counteracting it. This is accomplished, of course, by making the temperature of the upper

part of the column higher than that of the lower part. With this end in view the crucible was suspended in most cases in a part of the furnace where the temperature increased upward. In one experiment recorded here this method of eliminating convection was replaced by a method involving the use of a bath of molten gold to obtain a uniform temperature. The platinum crucible containing the charge was protected from the molten gold by a tube of silica glass.[1] The silica glass was rather soft at the temperature of the experiments and required reinforcement by a tube of Marquardt porcelain. The one result obtained by this method showed no significant difference from a result obtained by the method of suspending the bare crucible in the furnace. There seemed, therefore, to be no reason for preferring the use of the gold bath and it was not carried farther.

To make assurance doubly sure in the way of minimizing possible convection, the charge was made small. This acts in three ways: to make a possible lateral difference of temperature small, to render difficult the initiation of convection currents, and to decrease the velocity of possible currents. The crucibles used were, therefore, 5 to 6 mm. in diameter and 10 to 20 mm. deep. In order to simplify the conditions of diffusion the crucibles were right circular cylinders without flare or rounded bottoms.

The temperature was kept constant partly by continual watching and regulation and partly by using current from a storage battery. In some cases an automatic regulator was used, designed by W. P. White of this Laboratory.

In each case the heavy material taken was diopside. It was first melted and then chilled to a firm cake of glass $\left(G\frac{20}{4}=2.854\right)$ in the bottom of the crucible. The lighter material put in the top was one of the plagioclases Ab_2An_1, Ab_1An_1, or Ab_1An_2 $\left(G\frac{20}{4}=\right.$ 2.483, 2.533, 2.591 respectively). The temperature was in all cases about 1500°, that is, it was above the melting temperature of both layers, so that the experiments deal with diffusion of one liquid silicate into another.

[1] The use of silica glass in this manner was suggested by J. C. Hostetter, formerly of this Laboratory, who also kindly worked the glass into the desired form.

The charge was raised quickly to the desired temperature by plunging it into a furnace already somewhat above that temperature. After holding it for the desired length of time the charge was rapidly cooled by removal from the furnace, and the composition of the glass at various levels in the crucible was determined by measuring its refractive index, the relation between composition and refractive index having been previously determined on mixtures of known composition.

It may appear that the drastic temperature differences that arise when the cold crucible is placed in the hot furnace would be bound to set up violent convection, but fortunately this is a matter that can easily be ascertained by running a blank test. For this purpose a charge was prepared in the ordinary manner and allowed to remain in the furnace only a few minutes, when it was removed and the distribution of composition determined. It was found that the distribution was as it should be after diffusion for a short period with no random variations such as would result from convection.

In earlier experiments an ordinary thick-walled platinum crucible was used and the charge was always badly shattered in cooling. When not too numerous, the fragments were fitted together to reconstruct the original charge and the composition determined at various levels by removing a little powder with a file and determining its refractive index by the immersion method. The error involved in the measurement of the distance from the top or the bottom of the charge was large, and the results were only rough approximations. In later experiments the crucible was made of platinum foil 0.03 mm. thick. On contracting, the glass pulls the weak walls of the crucible with it and remains unshattered. The platinum foil was then peeled off and the cylinder of glass was ground to a wedge whose edge was parallel to the axis of the cylinder. The faces of the wedge were polished and the refractive index of the glass at various points was determined on the goniometer by the method of minimum deviation. The exact distance of the points from the bottom of the cylinder was measured by means of the scale on the centering screws of the goniometer.

Though numerous measurements were made by the early, rougher method, only the few made by the later method will be described. It may be noted, however, that the earlier results, within their rather large limits of error, agree with the later.

In order to obtain a visual impression of the diffusion of material, such as is obtained when copper sulphate crystals are placed in the bottom of a vessel of water, the layer of diopside was in some cases colored by the addition of 1 per cent Fe_2O_3 which imparts to the glass a marked green color. In all cases the change of color was sensibly coincident with the change of refractive index; in a layer where the index fell rapidly the color faded rapidly; in cases where an upper layer was found unaffected in refraction it was likewise uncolored.[1]

In summary it may be stated that the mode of procedure was as follows: A layer of diopside was placed in the bottom of a crucible with a layer of plagioclase above it, diffusion was permitted at constant temperature (above the melting-temperature of both layers) for a definite period, the charge was quenched, and the composition determined at various depths by measuring the refractive index of the glass.

GRAPHICAL PRESENTATION OF RESULTS

The results (Table I) may be presented graphically as in

TABLE I

DIFFUSION RESULTS

EXPERIMENT NO. 24

Depth of diopside layer 1.8 mm. Total depth 9.5 mm. Time 48 hrs.

Distance from bottom (mm.)	0.3	1.3	2.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3
Vol. per cent diopside	39.5	39.6	38.1	31.5	18.0	7.5	4.8	3.5	1.1	0

Plagioclase of upper layer Ab_2An_1.

EXPERIMENT NO. 27

Depth of diopside layer 3.2 mm. Total depth 11.7 mm. Time 48 hrs.

Distance from bottom (mm.)	1.0	2.0	3.0	4.0	5.0	6.0	7.0	8.0	9.0	10.0	11.0
Vol. per cent diopside	52.0	51.6	50.7	49.6	46.5	33.7	9.7	2.4	0.7	0	0

Plagioclase of upper layer Ab_2An_1.

[1] This result might be taken as indicating that selective diffusion was unimportant, though not necessarily absolutely lacking.

EXPERIMENT NO. 21

Depth of diopside layer 3.9 mm. Total depth 9.1 mm. Time 22 hrs.

Distance from bottom (mm.)	0.2	1.0	2.0	4.0	5.0	6.0	7.0	8.0	8.8
Vol. per cent diopside	48.0	47.9	47.5	45.5	43.0	41.0	36.5	30.5	29.5

Plagioclase of upper layer Ab_1An_1.

Immersed in bath of molten gold.

EXPERIMENT NO. 37

Depth of diopside layer 7 mm. Total depth 10.2 mm. Time 17 hrs.

Distance from bottom (mm.)	0.3	1.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3	9.5
Vol. per cent diopside	73.0	72.8	72.3	70.5	70.7	68.5	66.5	64.2	63.6	63.4

Plagioclase of upper layer Ab_1An_1.

EXPERIMENT NO. 25

Depth of diopside layer 3.2 mm. Total depth 14.5 mm. Time 22.5 hrs.

Distance from bottom (mm.)	1.3	2.3	3.3	4.3	5.3	6.3	7.3	8.3	9.3	10.3	11.3	12.3	13.3	14.2
Vol. per cent diopside	42.5	38.3	36.3	29.8	24.4	21.3	20.1	17.8	17.2	15.0	14.6	9.5	9.0	7.3

Plagioclase of upper layer Ab_1An_2.

Figures 1–5, in which the ordinates represent height in millimeters above the bottom and the abscissae composition in units per cent of diopside. The initial condition will then be represented by two vertical lines indicating two uniform layers, one at 100 per cent diopside and of a length corresponding to the depth of the diopside layer, the other at 0 per cent diopside representing the layer of plagioclase above it and of appropriate length. These are joined by a horizontal line indicating instantaneous change of composition. The final condition will be represented by a curve on which any point represents the composition at the corresponding level. The slope at any point indicates the composition gradient at that point, the curve approaching more nearly to the horizontal the greater the composition gradient. The figures need little discussion since they present the results better than can be done in words.

THEORETICAL CONSIDERATIONS

As noted on an earlier page, concentrations in a mass undergoing diffusion may be calculated on the basis of Fick's law. In its application to the present case, and in the form most useful for

calculation of the concentration at any point in a diffusion cylinder, the equation[1] becomes

$$2c = c_0 \left[\frac{2}{\sqrt{\pi}} \int_{\frac{-l-x}{2\sqrt{kt}}}^{\frac{l-x}{2\sqrt{kt}}} e^{-\beta^2} d\beta + \frac{2}{\sqrt{\pi}} \int_{\frac{-l-(2m-x)}{2\sqrt{kt}}}^{\frac{l-(2m-x)}{2\sqrt{kt}}} e^{-\beta^2} d\beta + \text{etc.} \right] \quad \text{I}$$

where the term in brackets in the limits is successively x, $2m-x$, $2m+x$, $4m-x$, $4m+x$, etc., and where c is the volume concentration at any point at distance x from the base of the column, l is the thickness of the bottom layer of original uniform concentration c_0, m is the total length of column, t is the time elapsed, and k the constant of diffusivity. For the examples in hand the series is rapidly convergent. With the aid of this equation we may, then, calculate the concentration at various points after a certain period of time and for a certain value of k (or, more simply, for a certain value of the product kt) and draw a curve representing the theoretical distribution of concentration. Curves of this kind were drawn and it was found that in no case could a calculated curve be obtained that would coincide with the observed curve. Of the calculated curves a certain one was chosen and was plotted on each of the figures as a dotted curve. The theoretical curve chosen in each case was that which showed approximately the same concentration at the upper surface as that actually found. The curves therefore coincide at their upper ends, but at other depths wide divergence is shown between the full curve of actual concentration and the dotted curve of theoretical concentration. In all cases this divergence is of a systematic kind, the actual concentration showing a smaller gradient in the diopside-rich layers and a larger gradient in the diopside-poor layers than the theoretical concentration. This is shown particularly plainly in Figure 1, where the diopside-rich layers have reached practical uniformity while the upper layers show a very strong gradient. This uniform

[1] The equation is not so formidable as it appears, $\frac{2}{\sqrt{\pi}} \int_0^q e^{-\beta^2} d\beta$ being merely the probability integral whose value, for various values of q in the limits, can be looked up in the tables.

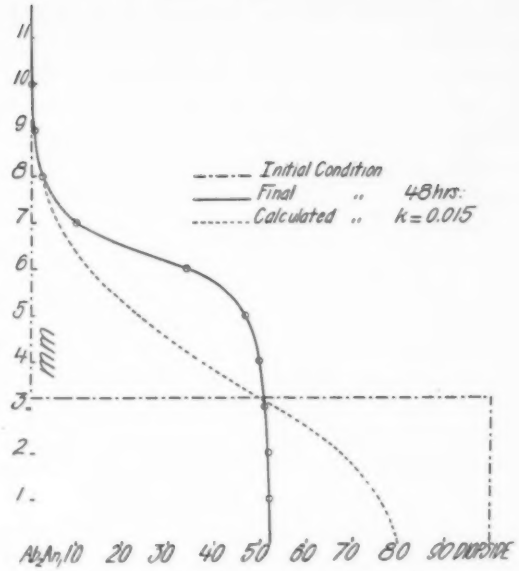

FIG. 1.—Diffusion experiment No. 27

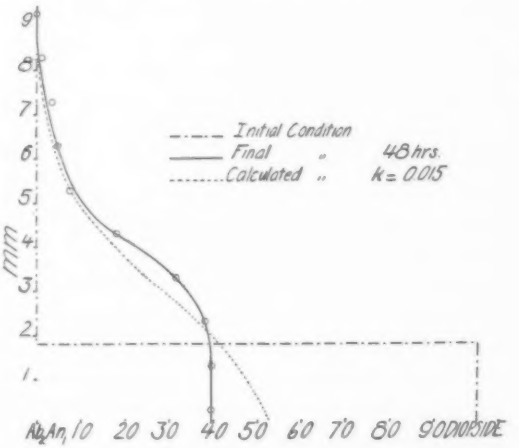

FIG. 2.—Diffusion experiment No. 24

divergence from theory may be explained by assuming that the coefficient of diffusivity is not a constant but is itself a function of concentration and is greater for diopside-rich mixtures than for those poor in diopside. In those experiments with aqueous solutions, where the highest degree of correspondence with theory is obtained, the solutions are always kept dilute so that the medium into which diffusion is taking place is sensibly constant. Under these conditions theoretical concentrations calculated on the basis of a constant value of the coefficient of diffusivity are in marked accord with observed values. In the present case, however, there is a continual and very important change in the nature of the diffusion medium as time progresses, and no constancy is to be expected in the value of the coefficient of diffusivity. There is no necessity, therefore, for regarding the results as showing divergence from Fick's law, the results being reconcilable with theory if it is assumed, as mentioned above, that the diffusivity is a function of concentration.

Einstein has developed for dilute solutions a theoretical relation between diffusivity and viscosity which makes them inversely proportional.[1] While experimental results do not entirely confirm his theory, they suggest its correctness if certain disturbing factors such as hydration could be evaluated.[2] At any rate, if we assume that the diffusivity is an inverse function of viscosity, we obtain a natural explanation of our experimental results. The viscosity of the diopside-rich mixtures is much less than that of the plagioclase-rich mixtures. The coefficient of diffusivity for the diopside-rich mixtures should be correspondingly greater, and this we have found to afford a natural explanation of the deviation from theoretical values calculated in the ordinary way. Moreover, as one would expect, this deviation is more marked for the plagioclase liquid Ab_2An_1 whose viscosity is very much greater than that of diopside liquid, and less marked for the plagioclase liquid Ab_1An_2 where the viscosity contrast is not so great, while liquid Ab_1An_1 occupies an intermediate position in this particular (cf. Figs. 1, 3, and 5). Qualitatively, then, the experimental results suggest

[1] *Ann. d. Physik*, Vol. XVII (1905), p. 549.

[2] L. W. Öholm, *Medd. K. Wettenskapsakad Nobelinstitut 2*, No. 26, p. 21.

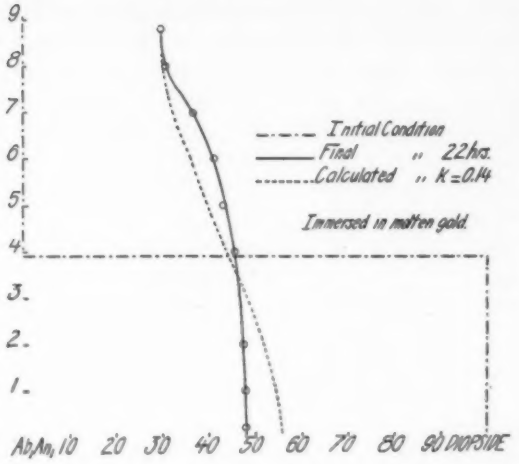

FIG. 3.—Diffusion experiment No. 21

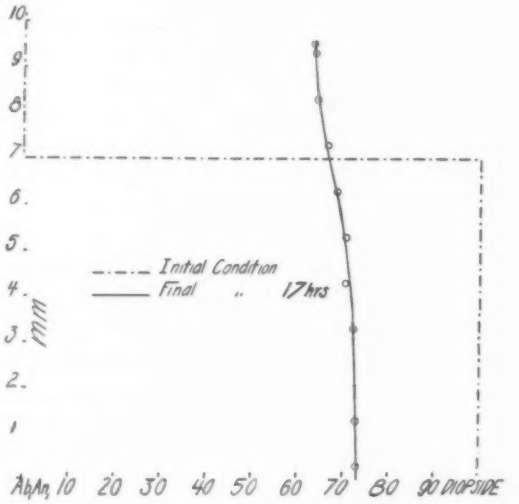

FIG. 4.—Diffusion experiment No. 37

agreement with theory if the variation of the coefficient of diffusivity is taken into account. It is possible, too, that from the observed values a definite quantitative relation between the coefficient of diffusivity and the concentration could be calculated, but the writer has not been able to find any attack upon a problem of this

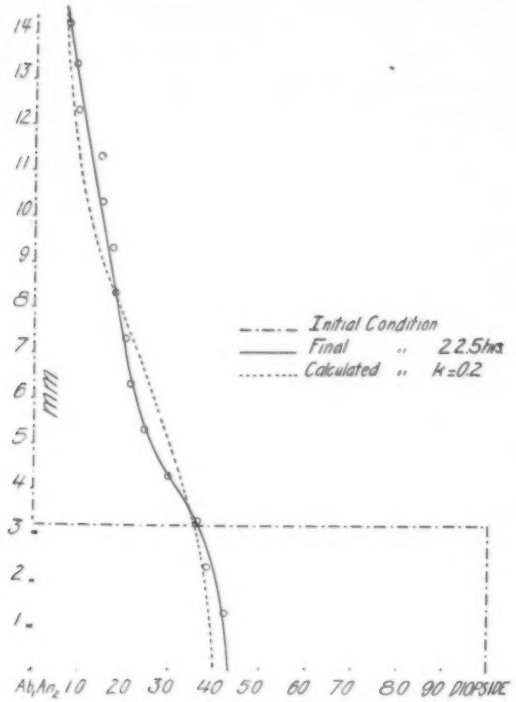

FIG. 5.—Diffusion experiment No. 25

kind in the various treatments of diffusion of concentration or of temperature. All of these adhere to a constant value of the coefficient of diffusivity. While in some respects it is highly desirable to check the present results more thoroughly along the lines indicated above, yet, for the purposes of the present paper, this is unnecessary. The results afford us very definite information as to the

magnitude of diffusion in silicate melts, which was the objective in mind when the work was undertaken.

VALUES OF DIFFUSIVITY

We cannot speak of a diffusivity "constant" in connection with the present results, but we may take the amount of diopside which penetrates to the surface layer as an indication of the average diffusivity of diopside in a liquid mixture of diopside and plagioclase. In this sense we find the "average diffusivity" of diopside in Ab_2An_1 mixture from Experiment No. 24 (Fig. 1), $k=0.015$ in cm.^{2} per day, in Ab_1An_1 mixture from Experiment No. 21 (Fig. 3), $k=0.14$, and in Ab_1An_2 mixture from Experiment No. 25 (Fig. 5), $k=0.2$. We therefore observe a progressive increase in the value of k with increase in the amount of anorthite in the diffusion mixture. Since it is well known that plagioclases rich in anorthite afford less viscous liquids than those rich in albite, we have further evidence of the increase of rate of diffusion with decrease of viscosity. If we compare the results of Experiment No. 21 (Fig. 3) and 37 (Fig. 4), we find again an increase in the value of the diffusivity, being $k=0.14$, and $k=0.3$, respectively.[1] In both these examples, however, the plagioclase Ab_1An_1 was used, the difference being that an increasing proportion of diopside was added. As a consequence, a higher value of the diffusivity is found for the mixture which was richer in diopside and therefore of lower viscosity. All of the results are therefore consistent with the assumption that the diffusivity varies inversely with the viscosity, which is in turn dependent on composition.

By way of comparison of the diffusivities here found with measured values of diffusivities for other substances, it may be noted that for common salt diffusing in water $k=1$ at 15°C., for gold in molten lead at 492°C., $k=3$, for solid gold in solid lead at 150°C., $k=0.0043$ in the same units as those used above. The

[1] In Figure 4 no calculated curve is shown for the reason that the calculated curve for $k=0.3$ sensibly coincides with the observed curve. Since the final result will be uniformity in all cases, whether the diffusivity varies with composition or not, a close approach of the calculated curve to the observed curve is to be expected in cases where diffusion is far advanced (Fig. 4). On the other hand, where the upper layers have not yet been affected, the greatest divergence, between observed and calculated values is to be expected (Figs. 1 and 2).

diffusivities obtained for the silicates are, therefore, much smaller than those of salts in solution and those of molten metals, but much greater than those of solid metals. Some of the higher values obtained for the silicates are comparable with those of certain relatively viscous organic liquids. The diffusivity constant of glycerine in propyl alcohol at 17°C. is, for example, approximately 0.2.

APPLICATION OF RESULTS

It is not at all likely that the diffusivities of substances in mutual solution in rock magmas can be significantly greater than those determined for the plagioclase-diopside mixtures, and in many viscous magmas they would no doubt be considerably less. For the purpose of applying the results to diffusion problems in petrogenesis a value has been taken very close to the highest, viz., 0.25, which is at the same time convenient in calculations. The Soret action is one of the diffusion phenomena that has been considered of possible importance in magmas. It has been found in the laboratory that if a tube containing a solution is heated at one end and cooled at the other there is usually a concentration of the solute toward the cold end which depends upon the difference of temperature, the relative concentrations being, for many cases, inversely as the absolute temperatures. In cooling magmas the margin must be regarded as having a lower temperature than the interior, and there should presumably be a tendency toward a greater concentration of some substance or substances at the cooler margin. This introduces the possibility of composition differences in different parts of an entirely liquid magma, the differences being brought about by diffusion. If cooled entirely by conduction, the temperature of a magma brought into contact with cold country rock should at the border quickly assume a value midway between that of the magma and that of the country rock. For a long period thereafter cooling at the margin is very slow (see Fig. 6).[1] We may imagine that the original temperatures of the magma and that of the surrounding rocks are such that during this long period of maintenance midway between them the magma is still above its temperature of beginning of crystallization. Here we would have

[1] See also Lane, *Ann. Rept. Geol. Surv. Michigan for 1909*, Fig. 18, p. 152.

the most favorable case conceivable for the establishment of a composition gradient as a result of a temperature gradient according to the Soret principle. These conditions would evidently obtain either when the temperature of the magma was very much above the crystallization temperature, or when that of the surrounding rock was not very much below it, the latter being the more likely case. The magma intruded into hot surroundings, perhaps into a cognate intrusive not yet cooled, is, therefore, the most favorable

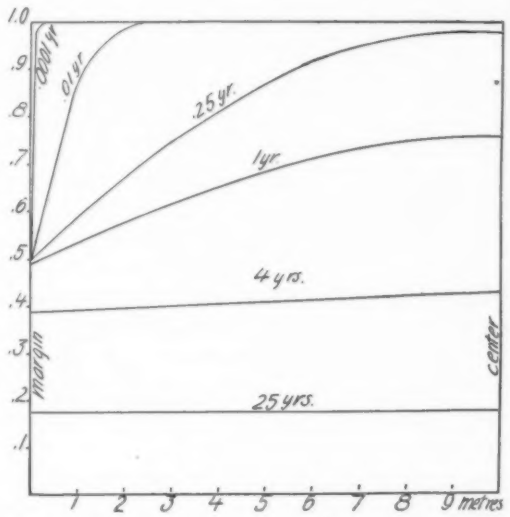

FIG. 6.—Curves of cooling of an intrusive igneous sheet 20 m. thick

subject for the working of the Soret action. Yet when we realize that the diffusivity of mass is, according to our determinations, from 10,000 to 100,000 times smaller than the diffusivity of temperature in rocks, it is apparent that the temperature of any igneous body will fall too rapidly to allow sufficient time for the Soret phenomenon to manifest itself.

This statement may perhaps be more readily appreciated if the Soret action is stated more definitely as a diffusion problem. In order to do so we may assume that the osmotic pressure is

proportional to the absolute temperature and that for this reason diffusion takes place until the concentration is inversely proportional to the absolute temperature. In other words, the effective concentration is, initially, inversely proportional to the absolute temperature, and diffusion takes place until the effective concentration is uniform. In applying these considerations to a cooling mass of rock, we may take for simplicity a tabular body. A solution of the problem of the cooling of such a body is given by the equation

$$\theta = \frac{\theta_0}{\sqrt{\pi}} \int_{\frac{-l-x}{2\sqrt{kt}}}^{\frac{l-x}{2\sqrt{kt}}} e^{-\beta^2} d\beta$$

where θ is the temperature at any point distant x from the margin, θ_0 the original temperature of the magma, the temperature of the wall rock being taken as zero and l is thickness of the intrusive.

If we take a tabular body of thickness 20 m. (i.e., 10 m. from center to margin) we may calculate the temperature in any plane at given distance from the margin at the end of any period of time. The results of such calculations are shown graphically in Figure 6, the curves representing the distribution of temperature at the end of various periods of time if the diffusivity is taken as 0.0118 in cm.^{2} per second. In this figure the temperature scale has no necessary absolute significance, 0 of the scale being merely the initial temperature of the surrounding rock and 1 of the scale being the initial temperature of the magma. It will be noted that at the end of one year the temperature at the margin is about halfway between the initial temperature of the surrounding rocks and that of the magma, while the temperature at the center is much higher. If it is supposed that the whole mass is still above its crystallization temperature, then the Soret action should be operative, that is, the effective concentration at any point should be proportional to the absolute temperature, and diffusion should take place until the effective concentration was uniform. Partly for simplicity and partly for the sake of obtaining an especially marked Soret effect we shall assume that the temperature scale of Figure 6 represents

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absolute temperature, i.e., 0 of the scale is 0° absolute and 1 of the scale is 1000° absolute. After one year the temperatures at the margin and at the center are 490° and 760° respectively. Then the effective concentration at the center should be $\frac{76}{49}$ times that at the margin and diffusion should take place until the effective concentration is uniform, or until the real concentration at the center is $\frac{49}{76}$ times that at the margin. The problem is to find how long a time it would require for this diffusion to take place. Infinite time would, of course, be required to allow the process to go to completion, but we may find the time needed to give any assigned approach to this condition.

As a first step we may calculate the time necessary for the acquirement, from any arbitrary initial condition, of a concentration gradient represented by the curve showing the thermal gradient at the end of one year. This may be done by assuming a condition analogous to our experiments, viz., that all the material was first concentrated in a meter layer and by diffusion had acquired the gradient referred to. With the aid of equation (I) we find this to be very nearly true when $\sqrt{kt}$ in the limits of the integral has the value 500. If we take k as having a value close to the highest found in any of the experimental determinations, viz., 0.25, then $t = 10^6$ days. But this is not the time we wish to know; it is that required to go on from this condition to practical uniformity. Again we find from the equation that practical uniformity (1:0.996) is obtained from the arbitrary initial condition when $\sqrt{kt} = 1000$ or when $t_1 = 4 \times 10^6$ days. From this we get the desired time $t_1 - t = 3 \times 10^6$ days, or nearly 10,000 years. This shows that it would need about 10,000 years to obtain nearly the full theoretical Soret effect required by the curve of temperature distribution after one year in a mass of the dimensions chosen. In the meantime, as shown by the curves of Figure 6, the whole mass would have cooled to the temperature of the surrounding rocks. Even if we imagine it to be still above its crystallization temperature at the end of four years, it will be noted that most of the temperature gradient has been destroyed at that time so that

there is no reason for the action continuing even for four years. It should be observed that, although the times have been computed for a body of a definite size, the solution is really of a general nature, for if the body were n times as thick, the time on the one-year curve would be changed to n^2 years and the time required for the Soret phenomenon would be n^2 times as large. Nothing is gained, therefore, for the Soret effect by making the body larger or smaller.

In assuming that the scale of Figure 6 represents absolute temperature, we have, of course, taken an impossible condition for any body of rock. This would mean that the surrounding rocks were initially at 0° absolute and the magma had not yet begun to crystallize at 490° absolute = 217°C. The assumption was made on account of the convenience of referring both concentration and temperature gradients to the same curve, but even if we make reasonable assumptions as to the temperature of the magma and of wall rock, our conclusion will not be affected. We may even assume that the Soret effect for silicates is many times that deduced from the theoretical (absolute temperature) relation, yet the outstanding fact remains that the time required to produce a significant amount of concentration of material by diffusion is enormously greater than that required for the mass to cool off.

It may be noted that in speaking of a concentration toward the cool margin no mention has been made of what is concentrated. The reason is, of course, that it is not known. Ordinarily it is stated that the solute is concentrated toward the margin, but no distinction of solvent and solute can be applied to magmas. In conclusion, then, it may be stated that no concentration of any substance toward the cool margin could occur in appreciable amount in the time available for such action in a cooling mass of completely molten rock.

DIFFUSION TOWARD MARGIN DURING CRYSTALLIZATION

There is, however, another case of diffusion of material toward the cool boundary for which we know the nature of the material that should move in that direction. Harker lays stress on the fact that in any cooling mass of magma there should be a time

when crystallization of an early-formed mineral A takes place only near the border, the rest of the mass being still above the temperature of crystallization. The greater concentration of the substance A in the interior of the mass where none has yet precipitated should constitute a driving force tending to cause that material to diffuse toward the margin. This case may be stated fairly simply as a definite diffusion problem. When the temperature of a thin layer at the margin has fallen to such a value that a certain fraction of the amount of substance A in that thin layer has precipitated, then the magma in this layer is saturated with A. If this condition is maintained for an infinite period of time, the whole body of magma will eventually acquire the same concentration in A as this marginal saturated solution[1] and all of substance A in excess of this concentration will be precipitated at the margin. If we assume the contact surface plane as in a tabular mass, this is essentially the same problem as the heat-conduction problem involved in the cooling of a sheet of metal one face of which is kept at constant temperature. A solution of the problem is given by the equation:

$$c = \frac{2c_0}{\sqrt{\pi}} \int_0^{\frac{x}{2\sqrt{kt}}} e^{-\beta^2} d\beta$$

which gives the concentration c in terms of the original concentration c_0 at any point at distance x from the margin after the time t, the concentration of the marginal saturated solution being taken = 0.

Values of c for various values of x and t have been calculated for a diffusivity 0.25 in cm.^{2} per day, approximately the highest experimental value, and are plotted as concentration curves in Figure 7. The figure shows that after two-thirds of a year the precipitating effect has been felt for a distance of 0.33 m. from the margin, all the rest of the magma being entirely unaffected. After sixty-four years the precipitating effect has been felt for a

[1] Neglecting the Soret effect.

distance of 3.3 m. while all the rest of the magma is unaffected, and similarly for the other concentration curves.

We can, moreover, determine from the figure the amount of material that would be precipitated at the margin at the end of any period of time. At the end of infinite time the concentration curve would correspond with the axis of abscissae, that is, the concentration at all points is zero of the scale, or equal to the concentration of marginal saturated solution. The whole area

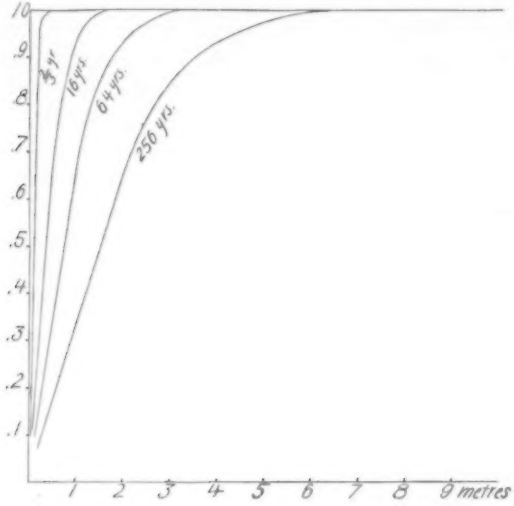

FIG. 7.—Curves of concentration in an igneous mass showing diffusion of material toward a cool boundary.

of the rectangle between the extreme ordinates of the figure represents, therefore, on a certain arbitrary scale the amount of material which would be removed in bringing the whole magma to the concentration of the marginal saturated solution. On the same scale the area lying above and to the left of the concentration curve for any time represents the amount of material removed during that time. We may take a specific case and imagine that the mineral precipitated at the margin is, say, amphibole, which occurs in the magma to the extent of 20 per cent, and that the

cooling of the marginal layer had proceeded until one-fourth of the amphibole contained in that layer (or 5 per cent of the layer) had been precipitated. If this condition were maintained the precipitation of amphibole at the margin would proceed by diffusion from the parts not yet cooled below the temperature of precipitation of amphibole. At the end of two-thirds of a year a proportion of the excess amphibole would be precipitated equal to the area between the one-year curve and the axis of ordinates. Regarding this area as a triangle of base sensibly 0.3 m. and assigning any arbitrary total thickness x to the mass, then the thickness of the amphibole deposit in meters would be

$$\frac{0.3}{2} \cdot \frac{1}{x} \cdot \frac{5}{100} x.$$

That is, a deposit 0.0075 m. or $\frac{3}{4}$ cm. thick would be formed on the margin in two-thirds of a year. Its thickness is independent of the total thickness of the mass of magma and all of it would be derived from a layer of magma 33 cm. thick. After sixteen years the deposit would be about 2 cm. thick, all coming from a layer of magma about 1.5 m. wide. After two hundred and fifty-six years the deposit would be about 8 cm. thick, all from a border portion less than 7 m. wide.

It is apparent that the possible results of diffusion after the manner postulated are exceedingly small. A mass of magma large enough to remain in the necessary condition for two hundred and fifty-six years would have a border phase 8 cm. thick. By necessary condition is meant that the margin should be cooled within its crystallization range and the main portion of the mass be not yet so cooled. The indications are that the mass would require to be at least 300 m. thick and be intruded under special conditions of temperature of magma and of wall rock, and the border phase would then be insignificant. Even by making more liberal assumptions as to the amount of chilling at the margin, say, a chilling sufficient to precipitate 25 per cent of the magma solution, the possible border phase would be increased in magnitude only five times. Moreover, as one increases the extent of marginal chilling, a stage is soon reached where so much precipitation takes

place in the marginal phase that no diffusion into that region can occur. One then arrives at a method of formation of a border phase that has been suggested by Daly, who regards the border phase as a chilled phase having the composition of the original magma.[1]

FORMATION OF REACTION RIMS

We have seen in the foregoing that the movement of large quantities of material through long distances by diffusion in a magma cannot be credited when the relatively rapid rate at which the magma must cool is considered. On the other hand, diffusion through short distances is to be expected, and such phenomena as the formation of reaction rims about foreign inclusions are readily to be attributed to diffusion. At the same time it should be noted that a rather wide reaction border will require a very considerable period of time for its formation if diffusion alone is active. Figure 7 enables one to form an idea of the period of time required for the diffusion of material from an inclusion to various distances in the surrounding medium if the scale of concentrations is reversed, that is, if zero is placed at the top and one at the bottom. The solution is by no means a rigid one for a small inclusion, but for a large slablike inclusion is sufficiently good to enable one to draw general conclusions. The figure shows that after sixty-four years the effect of the inclusion is barely felt for about 3 m. and is strongly felt (one-half saturation) for not more than 1 m. These considerations suggest that the formation of reaction rims up to 2 m. thick, such as those described by Ussing, about inclusions of quartzite in augite syenite at Kangerdluarsuk would require a period of time of the order of magnitude of one hundred years if diffusion alone were operative.[2]

The growth of crystals is itself largely dependent upon diffusion, but no quantitative estimate of the rate of growth is possible without some knowledge of the concentration gradient along which flow of material takes place, that is of the degree of supersaturation possible in the liquid interstitial to the crystals. The

[1] *Igneous Rocks and Their Origin*, p. 237.

[2] *Geology of the Country about Julianehaab, Greenland*, p. 362.

fact that rocks are normally millimeter-grained rather than centimeter- or meter-grained even in large masses is, however, a tribute to the slowness of diffusion in magmas. The fact that certain conclusions are reached above on the assumption that diffusion acts alone should not be taken as indicating that the writer believes that no other processes could occur. To account for the extremely coarse grain of many pegmatites, for example, it seems necessary to assume circulation of solutions, and in many other cases cited circulation (convection) would be inevitable.

SUMMARY

The rate of diffusion in certain silicate melts has been determined experimentally by permitting diffusion against gravity of a heavy liquid into a lighter liquid. The concentration curves found are not coincident with any theoretical curves calculated on the basis of a constant value of the diffusivity, but can be interpreted on the assumption that the diffusivity varies with concentration and is less for concentrations corresponding to more viscous liquids than for those corresponding to less viscous liquids. Taking as representative of the "average diffusivity" the amount of material which penetrates into the upper layer, the following values of the average diffusivity (k) were found: for diopside into Ab_2An_1, $k=0.015$; for diopside into Ab_1An_1, $k=0.14$ to 0.3, depending on the proportions; and for diopside into Ab_1An_2, $k=0.2$, all in cm.^{2} per day.

The value 0.25 (close to the maximum experimental value) is taken as probably representing a fair estimate of diffusivity in magmas, and with this as a basis it is shown that such phenomena as the formation of border phases about large bodies of igneous rock by diffusion cannot be considered possible in the time available for such action in a cooling magma. On the other hand, the formation of reaction rims about inclusions may be attributed to diffusion, though for very wide rims a considerable period of time will be required.

THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION AND MAGMATIC DIFFERENTIATION OF IGNEOUS ROCKS

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TRONDHJEM, NORWAY

INTRODUCTION

In a later paper of this treatise (I–IV) are given the physico-chemical laws which govern the *crystallization* of igneous rocks. Subsequently, it will be shown that the same laws can be applied to the explanation of the *chemical composition* of igneous rocks, and consequently also of *magmatic differentiation*.

As I shall often refer to my earlier publications on the problems here discussed, I give a list of those of most importance:

"Studier over slagger," *Svenska Vet.-Akad. Handl.*, 1884. (Stockholm, 1885.)

"Beiträge zur Kenntnis der Gesetze der Mineralbildung in Schmelzmassen und Ergussgesteinen," *Archiv for Mathem. og Naturv.*, Vols. 13 and 14. (Kristiania, 1888–90.)

"Die Silikatschmelzlösungen," I and II. *Kristiania Videnskabs-Selskab*, 1903, 1904.

"Physikalisch-chemische Gesetze der Krystallisationsfolge in Eruptivgesteinen," *Tschermaks min. und petrogr. Mitt.*, Vols. XXIV (1905), XXV (1906), and XXVII (1908).

"Über anchi-monomineralische und anchi-eutektische Eruptivgesteine," *Kristiania Vid.-Selsk.*, 1908.

"On Labradorite-Norite with Porphyritic Labradorite-Crystals: a Contribution to the Study of the "Gabbroidal Eutectic," *Quart. Jour. Geol. Soc.*, 1909.

"Über das Spinell: Magnetit-Eutektikum," *Kristiania Vid.-Selsk.*, 1910.

"Die Sulfid: Silikatschmelzlösungen" (a review, 97 pages), *Norsk Geologisk Tidsskrift* (Kristiania), IV (1917).

"Die Sulfid: Silikatschmelzlösungen. Part I. Die Sulfidschmelzen und die Sulfid: Silikatschmelzen." *Kristiania Vid.-Selsk.*, 1919. Later will appear Part II. Die Nickel-Magnetkies-Lagerstätten.

For geological surveying, chemical analysis, photographs, etc., for this publication I have had contribution from Den Tekniske Hoiskoles Fond (The Foundation of the Technical University of Norway).

ERRATUM

In the article by J. H. L. Vogt, p. 318, in the first line, delete the words "a later paper of."

I

REVIEW OF THE PHYSICAL CHEMISTRY OF THE CRYSTALLIZATION OF IGNEOUS MAGMAS

In the examination of these laws two different methods can be used: (*a*) the *synthetic*, in which there is an opportunity for precision-determinations, especially of temperature. Previous investigations on the crystallization of silicates from melts have been, nearly without exception, conducted at the pressure of one atmosphere; (*b*) the *analytic*, mainly based on the study of the structure of the rocks. In this manner we may examine the sequence of crystallization, and so also the "individualization-fields" of the minerals, further the mix-crystal systems, the chemical composition of eutectic intergrowths, etc.—all under the physical conditions, especially with regard to pressure and time, present during the solidification of the different igneous rocks.

The synthetic method forms the important base. The analytic method gives us, in particular, information as to the extent to which the results of investigations conducted chiefly at atmospheric pressure and during short periods of time, can be transferred to apply to the physical conditions under which the crystallization of magmas took place.

The two methods, therefore, go hand in hand and complete each other.

REMARKS ON THE STRUCTURAL CRITERIA FOR THE SEQUENCE OF CRYSTALLIZATION

The sequence of crystallization in igneous rocks may usually be determined by the complete, partial, or wanting idiomorphism of the minerals, by the inclusions, by deposition on a solid body (*Fixkörper-Absatz*), by "together-swimming" structure (*synneusis-struktur*, see below), by law-governed intergrowths, etc.

The complete *idiomorphism* of a primary mineral A against all the other minerals shows that its crystallization was finished before the commencement of the solidification of the others. From the partial idiomorphism of the primary mineral A against the primary mineral B whose idiomorphism is wanting, we can infer that the crystallization of A had commenced at an earlier stage than the crystallization of B. But we must not draw the more extensive

conclusion that A in its entirety had crystallized before the commencement of the crystallization of B. As to the conceptions allotriomorphism, hypidiomorphism, and panidiomorphism, we refer to the petrographic textbooks.

Inclusions of an idiomorphic, primary crystal A in B implies that A had crystallized earlier than the surrounding parts of B. But if A only appears in the exterior zone of B, the interior part of B may have crystallized earlier than A. And even if inclusions of idiomorphic crystals of A appear evenly distributed over the whole of B, also in the kernel of B, it may be that part of A also may have crystallized at a later stage. As an example, idiomorphic crystals of apatite, as is known, in many cases appear in the oldest silicates and in the ore minerals, indicating that the apatite crystallized before the commencement of the solidification of the iron ore and the silicates. But I warn against the conclusion, which is often drawn, that the apatite in its entirety crystallized during the earlier stage.

Further, it must be taken into consideration that small portions of the mother-liquid occasionally may be inclosed or included in a mineral during its growth. As example we refer to the well-known zonally arranged glass-inclusions in leucite, sanidine, etc., in many dyke and effusive rocks. If corresponding magma-inclusions occur in deep-seated rocks, a complete crystallization of this material will take place. Thus the result will be inclusions in the host of a later-crystallized mineral.

Inclusions of mineral A in mineral B may furthermore be due to the fact that A originally, at high temperature, occurred as a solid solution in B, and that afterward, owing to reduced solubility by decreasing temperature, A separated from the solid solution. As an example we refer to the well-known inclusions of perthitic albite or albite-oligoclase in the microcline of granites, etc. The microcline (or orthoclase) dissolved about 28 per cent Ab+An, the greater part of which later separated during refrigeration. Further may be mentioned the secretion of lamellae of monoclinic pyroxene in orthorhombic pyroxene[1] and conversely also of ortho-

[1] Cf. the general account in my publication in *Tscherm. min. u. petrogr. Mitt.*, Vol. XXIV (1905), pp. 537–42.

rhombic pyroxene in monoclinic.[1] In the same manner the well-known microscopic inclusions, often with idiomorphic contour, of titanic iron ore in hypersthene, diallage, and plagioclase of gabbros may be explained.[2] The latter inclusions were often interpreted by earlier investigators as older than the host-mineral, but in their present form they must be explained as later secretions from an originally solid solution. As may be understood from this account concerning the inclusions we must take into critical consideration a great number of momentums in determining the successive age of the minerals.

When a substance is segregated from a solution, it often, as is well known, adheres to a solid already present (solid body or *Fixkörper*). The result of this is the *deposition on a solid body* (*Fixkörper-Absatz*), which is also very important in the solidification of igneous rocks. We may here, for instance, refer to Figure 34, illustrating the deposition of spinel on pyrite; to Figure 33, illustrating the deposition of titanomagnetite on olivine; and to Figure 35, where in one place pyrite has been deposited on apatite while in another apatite has been deposited on pyrite.

The individuals of a mineral, segregated from a magma at an early stage, frequently swam together to assemblings or aggregates, the result of which is a structure, for which I propose the term *together-swimming structure* or *synneusis structure*.[3]

This together-swimming may occur very rapidly. I refer to my publication "Die Sulfid: Silikatschmelzlösungen" (I, 1919), Figure 11, illustrating assemblages of octahedrons of magnetite in a bessemer-matte, consisting chiefly of Cu_2S, and to Figure 28, *a*, illustrating assemblages of small individuals of zincblende in a slag. The solidification period of the two molten masses just mentioned, respectively molten sulfide and molten slag, needed only a very short time, at most half an hour.

[1] W. Wahl, "Die Enstatitaugite," *Tscherm. min. u. petrogr. Mitt.*, Vol. XXVI (1906).

[2] In this connection we refer to a treatise by A. Johnsen, "Regelmässige Einlagerung von Eisenglanz in Cancrinit," *Centralbl. f. Min., Geol. und Pal.*, 1911, and by O. Andersen, "On Aventurine Feldspar," *Amer. Jour. Sci.*, Vol. XL (1915).

[3] Composed of *σύν, syn* = together and *νεῦσις, neusis* = swimming.

The phenomenon here discussed may, as to igneous rocks, be illustrated by Figure 1, representing a dunite from the Hestmandö-field in the northern part of Norway, with an average of only about 1 per cent chromite. In some parts of the thin section chromite is entirely or almost entirely lacking, but in other places we may find aggregates of ten to twenty small octahedrons of

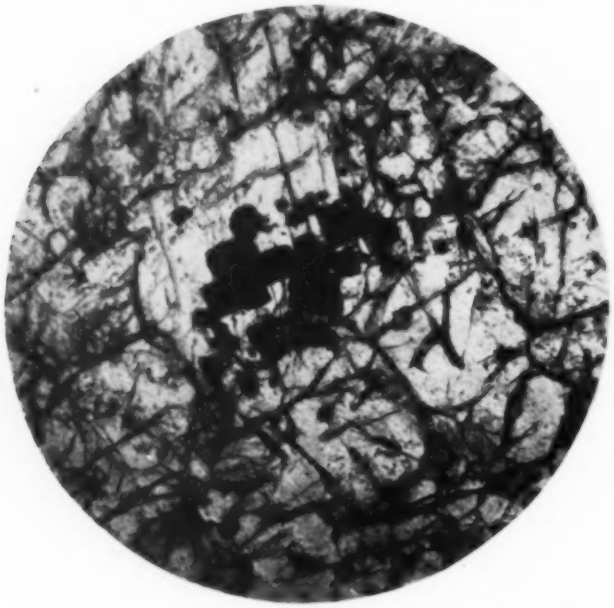

FIG. 1.—Dunite from the Hestmandö-field, northern Norway. Groups of octahedrons of chromite, illustrating "together-swimming" or synneusis structure. (Photo. 25:1.) (Black = chromite, white = olivine.)

chromite which, as well with reference to the idiomorphism against the olivine as with reference to the together-swimming structure, must have crystallized while the olivine was still in a molten condition. From this and other dunite rocks with a little chromite we may draw the conclusion that the chromite commenced crystallizing earlier than the olivine, when there was at least 1 or 0.5 per cent, perhaps only 0.33 per cent, chromite present. But

this does not exclude that the olivine may have commenced crystallizing earlier than the chromite when the latter only amounted, for example, to 0.1 or 0.05 per cent. The earlier silicate minerals, for example, olivine in olivine-rich gabbros, and hypersthene or diallage in hypersthene- or diallage-rich gabbros, also often show the together-swimming structure, as in Figures 10, 12, 13, 20 and 21, and 33.

The *relative commencement* of the solidification, especially of the minerals that commence crystallizing at a somewhat early stage, may often quite easily be decided by the structure. On the other hand, the allotriomorphism of a mineral C, against the minerals A and B, shows that C only commenced crystallizing after an often quite essential part of A and B had already solidified. Especially where the later mineral C is present in a small quantity, its allotriomorphism in connection with its appearance as *Zwischenklemmungsmasse* (or mesostasis) presents an easily recognizable criterion that it belongs to a very late stage of the crystallization. But at this late stage the minerals A and B will in many cases have continued forming. We refer to the explanation given in connection with Figures 17 and 18.

The *simultaneous crystallization* of two or more minerals may be manifested in various ways. With two simultaneously crystallizing minerals, each may grow until the individuals of A happen to collide with the individuals of B. Or some of the segregating mineral A may be deposed on the already solidified crystals of B, and some of the simultaneously segregating mineral B on the already solidified crystals of A.

In this manner we may observe crystals of plagioclase with quite good idiomorphic contour against the hypersthene or diallage and, further, crystals of hypersthene or diallage with quite good idiomorphic contour against plagioclase in the same thin section of an anchi-eutectic norite or gabbro. In the deep-seated rocks it may in such cases often be quite impossible to decide which of the two minerals first commenced crystallizing.

It may also often occur that the two minerals crystallize in an intimate intergrowth. In this manner simultaneously crystallizing

feldspar and quartz, as is well known, sometimes produces micropegmatitic or granophyric structure, as in graphic granite. Corresponding structure, which gives an evidence of a crystallization along a eutectic boundary curve, or exceptionally by a binary (or ternary or still more complex) eutectic, is also sometimes found with other minerals, for instance, between olivine and magnetite (see Fig. 28).

In many, possibly in most, cases the crystallization of the minerals A, B, C, etc., takes place in the following manner: Each mineral begins crystallizing at its proper stage, and continues to grow until the entire magma has solidified. As an extreme example we may choose apatite. This phosphate is only slightly soluble in silicate magmas at temperatures just above that at which the silicates commence to crystallize. If there is 0.20 per cent apatite present, the essential portion, perhaps 0.18, 0.19, or 0.195 per cent, has already crystallized before the silicate minerals have commenced segregating. But we may be pretty certain that a trifle phosphate, 0.02, 0.01, or perhaps only 0.005 per cent, still exists in solution at this stage and little by little solidifies later. It has, however, not been possible for me to substantiate this by observation with respect to the apatite; but I have been able to establish that spinel, when present only as 0.01 or at most 0.02 per cent, only commenced crystallizing after a great part of the silicate mineral A had solidified. (See Fig. 33 and the chapter on spinel.)

Fe_3O_4 and the different ferromagnesian silicates are only slightly soluble in acid—or granitic—magmas, and therefore commenced crystallizing at an early stage. We find, however, as is discussed below, a small remnant of magnetite and ferromagnesian silicate in the final product of the solidification. We may consequently draw the conclusion that the essential part of the magnetite and the ferromagnesian silicate was certainly solidified during the first stage of the crystallization, but that a little remnant stayed in the solution and was solidified later.

In a binary system, type IV, of two discontinuous mix-crystals—A, melting at relatively high temperature (for example, FeS_2,

at high pressure), and B, melting at relatively low temperature (for example, FeS)—idiomorphic crystals of B never appear in A, but, on the contrary, idiomorphic and usually somewhat resorbed crystals of A appear in B. Here the sequence of crystallization, without regard to the proportion of weight between A and B, is first A and later B, and this to be understood thus, that the crystallization of A was *completed* before that of B began. (See Fig. 36.)

In the usual silicate eruptives, most frequently consisting of a whole series of components, we may also meet corresponding crystallization of a certain mineral, completely solidified at a relatively early stage. See Figures 20 and 21, representing crystals of hypersthene imbedded in diallage.

Another case of crystallization *completed* at an early stage is illustrated by Figures 8 and 9 (and a theoretical explanation given below) with crystallization at the beginning stage of hypersthene, while the Fe-Mg silicate at a later stage entered into biotite.

In this treatise (Part I) we are only going to consider the solidification of the rocks (the transition from liquid to solid phase). We, however, also discuss the continued change of the minerals, which may be founded on the later crystallization of a substance originally in solid solution, and furthermore we are going to deal with the reactions which appear in the solid phase on the boundary between two minerals, and which are an immediate result of the cooling of the rocks after completed crystallization. We shall, however, not discuss the later changes, which are not a direct result of the solidification of the rocks, but are founded on exterior incidents, as, for instance, dynamo and contact metamorphosis, chemical actions, etc.

INTRODUCTORY REMARKS CONCERNING THE APPLICATION OF THE PHYSICO-CHEMICAL LAWS TO THE CRYSTALLIZATION OF MAGMAS

Magmas usually consist of a whole series of components, which entail a complication of the equilibrium existing in the magma and consequently also of the laws of the crystallization.

In many cases, however, an essential simplification of these complications takes place, as, according to H. E. Boeke,[1] the

[1] *Grundlagen der physikalisch-chemischen Petrographie* (1915), p. 104.

following assertion is applicable: "The saturation-boundary between binary mix-crystals as well as in general the equilibrium between two solid phases of a binary system do not change by contact with other phases and components, when these new entering components do not form solid solutions or stoechiometric compounds with the former solid phases." The proportion between An and Ab in the segregated plagioclase mix-crystal will in this manner be the same whether the crystallization takes place in a pure An+Ab melt or in a silicate melt (or magma) which besides plagioclase also delivers, for instance, magnetite, olivine, etc.

In order to investigate the laws of crystallization of the principal components of the magma, we may generally leave the components which are present only in subordinate quantity out of consideration, provided that the latter do not form solid solutions or enter into mix-crystal combinations with the principal components. We must, however, take into consideration that when A, B, and C form a ternary eutectic, and C only is present in small quantity, the simultaneous crystallization between A and B, along the eutectic boundary between A and B, will not be identical quantitatively with the composition of the binary eutectic between A and B. If C, however, is present in minimal quantity, the difference between the point in question on the eutectic boundary between A and B and the binary eutectic A:B will be so inconsiderable that it practically may be left out of consideration.

THE FELDSPARS, AB:AN, OR:AB+AN.

The binary system *Ab*:*An* (with melting-point Ab = 1100 ± 10° and An = 1550° ± 2°) belonging to mix-crystal type I has been studied in detail by N. L. Bowen[1] of the Geophysical Laboratory of the Carnegie Institution of Washington at the pressure of one atmosphere (and with chemically pure substances). We reprint Bowen's graphic exhibit as Figure 2, where the great horizontal difference between the liquidus and solidus curves is shown.

As an example we may mention that from a molten mass, Ab_1An_1, the first mix-crystal, separating at 1450° (without super-

[1] *Amer. Jour. Sc.*, Vol. XXXV (1913).

saturation), has a composition nearly exactly $Ab_{20}An_{80}$, or Ab_1An_4. Even before Bowen's investigation (1913), I had (see especially *Tscherm. Mitt.*, Vol. XXIV [1905]), on the basis of the well-known zonal structure of the plagioclases, and further on the basis of the proportion between Ab:An in the total plagioclase calculated from the analysis of the rocks and in the first segregated crystal, decided that the system Ab:An belongs to type I, and, furthermore, that in the crystallization of plagioclase in eruptive rocks, especially in

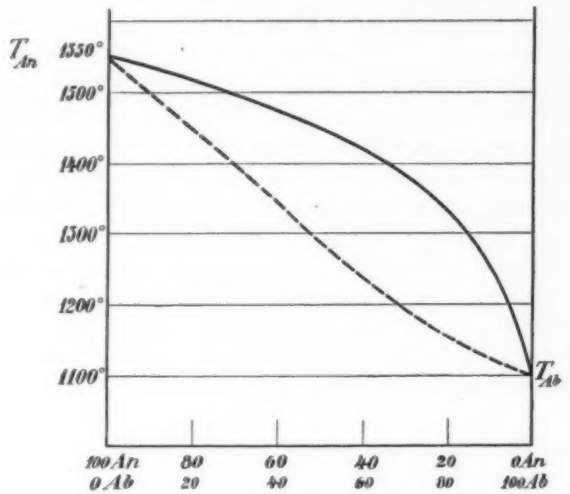

FIG. 2.—The melting-diagram, An:Ab, after Bowen

dike and effusive rocks, there is a very great horizontal difference between the liquidus and the solidus curves.

The binary diagram for Ab:An, at the pressure of one atmosphere and for 100 per cent Ab+An, may with unessential modifications of the horizontal difference between the two curves (see a following chapter) be transferred to magmas, crystallizing at high pressure, which besides Ab and An contain other components. The investigation of the mix-crystal system and the proportion between the liquidus and solidus curves for Ab:An may consequently also be applied to rocks, which besides plagioclase also

contain a great quantity, 40 per cent and still more, of foreign components, as quartz and Fe-Mg silicates. Petrographic experience proves that in such rocks we can detect no difference in the horizontal distance of the two curves, and this is in best accordance with the general law cited on page 326.

Or:Ab or Or:Ab+An.—Because of the extreme viscosity which characterizes melted $KAlSi_3O_8$ and $NaAlSi_3O_8$ (without or with only a small quantity of $CaAl_2Si_2O_8$), the synthetic study of the system Or:Ab or Ab+An is connected with exceptionally great difficulties. We may therefore here use the analytic method, cf. my earlier publication in *Tscherm. Mitt.*, Vol. XXIV (1905).

When Or is predominant, orthoclase first crystallizes, and when Ab+An is predominant, plagioclase first. The boundary ("individualization-boundary") is decided by the following method of investigation:

In a number of rocks, where the proportion Or:Ab:An in the entire rock was determined on the basis of the rock analysis, we find crystallized as No. I: *plagioclase* at 32, 32, 32.5, 32.5, 33.5, 34, 36, 36, 37.5, 39.5, 40, and 41 Or:Rest Ab_{+An}; *orthoclase* (or microcline) at 42, 43, 43.5, 43.5, 46, 47, 47, 50, 50, 50.5, 52, and 52.5 Or:Rest Ab_{+An}.[1] Ab_{+An} here represents Ab with a small but variable quantity of An, consequently albite, oligoclase, and andesine.

Between the two "individualization-fields" lies the boundary for orthoclase: albite, oligoclase, or andesine at about 0.4 Or:0.6 Ab or Ab_{+An}, perhaps nearest 0.42 Or:0.58 Ab_{+An}, applying to magmas at high pressure and with predominant Or+Ab+An with some mixture of other components. This boundary must be interpreted as an eutectic boundary curve.

When orthoclase (or microcline) crystallizes at *high* temperatures from granitic magmas, which besides predominant Or also contain a good deal of Ab_{+An}, there enters in the orthoclase up to about 30 per cent (or 28 per cent) of Ab_{+An}, which is partially separated by the cooling of the solid solution as albite or albite-oligoclase-

[1] Two uncertain determinations with 38 and 36 Or are here left out of consideration.

perthite. By the crystallization of acid plagioclase, the latter absorbs up to about 10 per cent (or 12 per cent) Or, which on cooling may give anthiperthite. Basic plagioclase (labradorite-bytownite) seems to absorb a somewhat smaller quantity of Or. Concerning this matter we refer to the chapter on the anorthosites in Part II.

By the above-mentioned crystallization in the acid magmas of Or and Ab (or Ab_{+An}) there result two minerals: orthoclase about 0.7 Or + 0.3 Ab (or Ab_{+An}), and acid plagioclase, about 0.9 Ab (or Ab_{+An}) + 0.1 Or.

Orthoclase and albite have almost exactly the same melting-point, and almost exactly the same atomic weight, probably also about the same latent melting-heat, etc. Granted a binary eutectic system (type V), the eutectic must lie almost exactly midway between 0.7 Or + 0.3 Ab and 0.9 Ab + 0.1 Or, consequently at 0.4 Or:0.6 Ab, just as we found above.

As An has a far higher melting-point than Or (and also than Ab), it is likely that the eutectic boundary between orthoclase and plagioclase with decreasing Or is characterized by increasing An in the plagioclase. In this manner we shall probably find the individualization-boundary between orthoclase and labradorite at 0.45—0.5 Or:0.55—0.5 Ab_{+An}.

This explanation is, however, not explicit, as we have not taken into consideration that in certain rocks or under certain conditions, the details of which we are not acquainted with, the two independent minerals, orthoclase and albite (or some other acid plagioclase), do not appear, but instead we find the mineral anorthoclase.

In my treatise in *Tscherm. Mitt.*, Vol. XXIV (1905) I pointed out the fact that a whole series of analyses of anorthoclase shows relations about 0.4 Or:0.6 Ab (or about 0.42 Or:0.58 Ab_{+An}) and I set forth the hypothesis that the anorthoclase might be defined as a microscopical or submicroscopical eutectic intergrowth of orthoclase (microcline) and albite (or some other acid plagioclase). This hypothesis is, however, quite dubious, and the physicochemical interpretation of the anorthoclase is still an open question.

QUARTZ (QU) AND FELDSPARS. QU:OR, QU:AB, QU:AN, QU:AB+AN, QU:OR:AB+AN

The binary system SiO_2:$CaAl_2Si_2O_8$ (*An*) has been examined by G. A. Rankin and Olaf Andersen[1] (at the Geophysical Laboratory, Washington) with the result: An, melting-point = 1550° ± 2°; Eutectic SiO_2:An = 52 per cent An:48 per cent SiO_2, melting-point 1353° ± 2°.

According to I. B. Ferguson and H. E. Merwin[2] (1918) the melting-point for tridymite is 1670 ± 10°C; for cristobalite 1710 ± 10°C. K. Endell and R. Rieke[3] (1912) decided for cristobalite 1685 ± 10°. N. L. Bowen[4] decided a somewhat lower temperature, and C. N. Fenner[5] thought he might fix the melting-point of cristobalite at 1625°, which, however, according to the latest precision-investigations, must be a little too low.

It is a matter of course that *Qu and Or*, as well as *Qu and Ab*, in the same manner as Qu and An, must form a binary eutectic. Because of their extreme viscosity the binary eutectics Qu:Or and Qu:Ab are not experimentally determined. We may therefore here use the analytic method.

We shall commence with *graphic granite*, the structure of which, as already established many years ago by W. C. Brögger,[6] is due to a simultaneous crystallization of quartz and feldspar. That is to say, the crystallization took place at a eutectic point or along a eutectic boundary curve.

Referring to my earlier publications[7] on the problem in question, we are going to give a collocation of all the hitherto published usable or at any rate somewhat usable analyses of graphic granite from pegmatitic granite dikes.

[1] The System Anorthite-Forsterite-Silica," *Amer. Jour. Sci.*, Vol. XXXIX (1915).

[2] *Amer. Jour. Sci.*, Vol. XLVI (1918).

[3] *Zeitschr. f. anorg. Chemie*, Vol. LXXIX (1913).

[4] *Amer. Jour. Sci.*, Vol. XXXVIII (1914).

[5] *Ibid.*, Vol. XXXVI (1913).

[6] *Geol. Fören. Förh.*, Vol. V (1881), and *Zeitschr. f. Kryst. Min.*, Vol. XVI (1890), I, pp. 148–59.

[7] "Silikatschmelzlös.," II (1904), and *Tscherm. Mitt.*, Vol. XXV (1906).

TABLE I

	No.	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	Total
	Microcline Graphic Granite										
Norway	1	74.04	14.40	Nil	Nil	Nil	0.33	2.01	9.36	Nil	100.18
	2	74.00	14.31	Nil	Nil	Nil	0.39	2.42	9.02	Nil	100.14
	3	73.82	14.44	Nil	Nil	Nil	0.35	2.45	8.90	Nil	99.96
	4	74.47	15.13	Nil	Nil	Nil	....	....	....	....	.....
	5	73.76	14.11	Nil	Nil	Nil	0.39	3.04	8.72	Nil	99.96
Sweden	6	72.41	14.51	0.30	....	Nil	0.13	2.19	10.00	0.28	99.91
	7	74.58	13.37	0.24	....	Nil	0.32	1.16	9.80	0.57	100.04
United States	8	73.86	13.75	0.26	....	Nil	Nil	2.10	9.00	0.24	99.24
	9	73.92	14.26	0.30	....	Nil	Nil	2.06	8.90	0.11	99.64
	10	72.76	15.44	($InAl_2O_3$)	....	Nil	0.19	2.35	9.28	0.15	100.20
	11	72.80	15.07	0.26	0.21	Nil	Nil	3.35	7.92	0.30	99.91
	Oligoclase Graphic Granite										
Norway	12	76.8	14.2	Nil	Nil	Nil	1.7	6.1	1.5	....	100.3
Sweden	13	76.67	14.20	0.14	....	0.04	2.67	5.33	0.52	0.48	100.04

EXPLANATION

Nos. 1–5 and 12, see "Silikatschmelzlös.," II. No. 1 from Arendal; Nos. 2–3 from Hitterö; No. 4 from Raade; No. 5 from Arendal is industrially pulverized graphic-granite, in which a trifle surplus feldspar is not excluded. In all these analyses a small loss (0.1–0.3 per cent) caused by ignition is deducted. The precipitate of Al_2O_3 was in all cases entirely white, and on account of this a special analysis of iron was not made. Some iron, less than 0.1 per cent Fe_2O_3, is, however, not excluded. No. 12, approximate analysis from Evje. Nos. 1 and 5 were analyzed by A. Grönningsæter and E. A. Dalset, assistants at the time. Nos. 2–4 and 6 were analyzed by students. K_2O,Na_2O and CaO in No. 4 have not been included on account of less accuracy.

Nos. 6, 7, and 13 from A. Bygdén, *Bull. of the Geol. Inst. of Upsala*, Vol. VII (1906); from Elfkarleö, Skarpö, and Ytterby.

Nos. 8–10 from Edson S. Bastin, *U.S. Geol. Surv. Bull. 420* (1910); from Topsham, Me., and Redford, N.Y. (No. 10, a little Fe_2O_3 [n.d.] in Al_2O_3.) No. 11 from Hiriart Hill, Cal., analyzed by W. T. Schaller, cited by H. S. Washington, "Chemical Analysis of Igneous Rocks, 1884–1913," *U.S. Geol. Surv. Prof. Paper 99* (1907). (Cited below as Wash.)

In the above table I have not included: Analysis of oligoclase graphic granite from the West Indies in Bygdén's treatise (Wash., p. 267, No. 5, with 68.12 per cent SiO_2), as this specimen is greatly decomposed, with considerable new-formed epidote, etc. This analysis can therefore not be used in the determination of the proportion of quartz and feldspar.

Analysis No. 4 in E. S. Bastin's treatise (Wash., p. 108, No. 4, with only 71.00 per cent SiO_2), as Bastin informs us of this specimen: "Some small

areas of pure feldspar were associated with the graphic granite in this specimen, so that the silica percentage shown in the analysis is lower than it would be for graphic granite alone." The same is probably also the case of an analysis, by A. W. Howitt, 1888, from Victoria (Wash., p. 112, No. 30).

With respect to the other analyses which are noted in Washington's Index as graphic granite we remark: "The graphic microgranite," page 73, No. 2 (with 3.74 per cent Fe_2O_3, 2.81 per cent FeO), represents a rock, and not graphic granite. The same also applies to No. 19, page 94 (with 0.73 per cent Fe_2O_3, 0.78 per cent FeO, 0.99 per cent MgO).

In the treatise, above cited, by Bygdén, as also in a treatise by H. E. Johansson,[1] an analysis by P. J. Holmquist[2] has been taken as an example of an albite-graphic granite, showing 77.32 per cent SiO_2, 0.34 TiO_2, 11.62 Al_2O_3, 1.57 Fe_2O_3, 0.69 FeO, 0.10 MnO, 0.62 CaO, 0.80 MgO, 0.99 K_2O, 5.81 Na_2O, 0.65 H_2O, *total* 100.51. This was computed by Holmquist as: 39.0 per cent Qu, 49.3 Ab, 4.6 Or, 1.9 An, also 2.4 chlorite, 2.3 magnetite, 0.8 titanite, 0.3 water, and a little calcite. The analysis is from a thin dyke in diabase (Rödö) and does not permit any exact determination of quartz:albite in graphic granite.

The precision-determination of the quantitative proportion of quartz and feldspar is complicated partly because of the inevitable errors in the analyses, of which more below, and partly because we cannot always be certain that the analyses represent absolutely pure intergrowths of the two minerals. In granite-pegmatite dikes we sometimes meet specimens of which one part consists of pure feldspar, free from quartz, and the other part of graphic granite, retaining the crystallographic orientation of the feldspar. That is to say, some feldspar first crystallized alone, and later, having reached an eutectic boundary curve, it continued its growth simultaneously with quartz. Sometimes we may find in the center of a large specimen of graphic granite small parts of pure feldspar. In such cases, the analyses of course cannot be used for precision-determinations of the relative proportions of quartz and feldspar. We may here refer to Bastin's remarks concerning his analysis No. 4.

On the basis of the analysis, we are going to calculate the quantitative proportions of quartz and feldspar in our microcline-graphic granites, according to the following methods:

[1] *Geol. Fören. Förh.*, Vol. XXVII (1905).

[2] *Sveriges Geol. Unders.*, C. 181 (1899).

a) Originating from K_2O, Na_2O, and CaO we calculate the quantity $KAlSi_3O_8$ (Or), $NaAlSi_3O_8$ (Ab), and $CaAl_2Si_2O_8$ (An), and when this is done the sum of feldspar is deducted from the analytically determined sum of SiO_2, Al_2O_3, K_2O, Na_2O, and CaO. The difference is quartz. This method contains a very great source of error, as an error in the determination of K_2O, Na_2O, or CaO in the calculation of the amount of feldspar will be doubled respectively 6, 8.5, or 5 times. To this may be added the inaccuracy in the SiO_2 determination.

b) Originating from the proportion $K_2O:Na_2O:CaO$ we calculate the percentage of SiO_2 in the feldspar (ex. 65.28 per cent in No. 1) and then we calculate the proportion between quartz (*n*) and feldspar (1−*n*), ex. for No. 1.: *n*. 100+(1−*n*). 65.28=74.04. A relative error in the determination of K_2O, Na_2O, or CaO will in this manner be eliminated. But an error in the determination of SiO_2 will be doubled nearly three times.

We have a control of the calculation, in method (*a*), in the calculated percentage of Al_2O_3 in the feldspar, compared with the percentage of Al_2O_3 found in the analysis. As an example, No. 5 shows 1.06 per cent too much Al_2O_3 in the calculation, consequently also too much feldspar, and so too little quartz. The calculated 21.58 per cent quartz must consequently be increased. On the other hand, No. 9 shows 1.12 per cent too little Al_2O_3 in the calculation, consequently too little feldspar or too much quartz.

With combined consideration of both methods of calculation, and of the sum of the analysis minus ignition and Fe_2O_3, we have under C written the probable percentage of the quartz. The rest is feldspar.

TABLE II

No.	A					B	C
	Qu	Or	Ab	An	Al_2O_3	Qu	Qu
1.....	25.60	55.51	17.05	1.64	−0.34	25.23	**25.4**
2.....	24.34	53.50	20.52	1.94	+0.16	24.90	**24.6**
3.....	24.54	52.78	20.78	1.74	−0.12	24.25	**24.4**
5.....	21.58	51.72	25.78	1.94	+1.06	23.52	**23.5(+?)**
6.....	20.55	59.84	18.57	0.65	+0.28	19.28	**20.3(+?)**
7.....	29.43	58.12	9.84	1.59	−0.25	27.53	**28.5(+?)**
8.....	27.01	53.38	17.82	.....	−0.53	23.57	**25.3**
9.....	27.32	53.32	17.47	.....	−1.12	23.67	**25.5**
10.....	22.94	55.04	19.94	0.95	−1.18	20.79	**21.9 (?)**
11.....	22.79	46.97	28.41	.....	−0.97	19.19	**21.0 (?)**

The specimen No. 6 is an erratic block (with 0.30 per cent Fe_2O_3 and 0.28 per cent ignition). Here a minimal, hardly perceptible, decomposition is not excluded, so the determination 20.3 per cent quartz (or according to Bygdén 20.81 per cent) is probably too low. For No. 11 I have no supply of the original literature. The analysis shows 0.26 per cent Fe_2O_3, 0.21 FeO, and 0.30 ignition, and to the calculated result is added an interrogation point.

Arthur Holmes[1] has used another method in order to determine the quantitative proportions of quartz and feldspar in graphic granite from dikes of granite-pegmatite (in Mozambique), viz., Rosiwal's planimetric method. In this manner, from different localities, he found quartz amounting to 27.9, 27.1, 26.3, 25.6, 25.3, and 24.2, average 26.1 per cent (calculated in percentage by weight) and rest, 73.9 per cent of microcline with the ordinary perthitic admixture of albite-oblioclase. Consequently we have the following determinations of the quantity of quartz in microcline graphic granite:

Calculated from the quantitative analyses: 28.5 (too high ?), 25.5, 25.4, 25.3, 24.6, 23.5 (?), 21.9 (too low ?), 21.0 (?), and 20.3 (too low ?).

By the Rosiwal method: 27.9, 27.1, 26.3, 25.6, 25.3 and 24.2 per cent of quartz.

The majority of these determinations are subject to great sources of error, which may amount to several per cent. If we take this into consideration, I think I am justified in drawing the conclusions that the proportions of quartz and feldspar in microcline graphic granite from dikes of pegmatite are subject only to small variations or are practically constant, and that we may fix the proportion pretty closely at:

26 per cent quartz:74 per cent microcline.[2]

The graphic granite in granite-pegmatite dikes crystallized at a relatively late stage, viz., after the essential part of the mica and, most frequently, also a part of the feldspar had solidified. Only a trifle mica was left at the time for the forming of the graphic granite. But in addition to this, besides the components of the feldspars and the quartz, there was surely some H_2O present, possibly partly connected with SiO_2 in forming a separate component (as H_2SiO_3 [?]). The graphic granite will thus have crystallized from a solution, which consisted predominantly of the components of feldspar and quartz, but also of a little mica and a small quantity of a component, as H_2O and H_2SiO_3 (?). The graphic granite has in this manner crystallized at a eutectic boundary

[1] *Quart. Jour. Geol. Soc. London*, Vol. LXXIV (1919), p. 77.

[2] In "Silikatschmelzlös.," II (1904), I gave the proportion 25.75:74.25, which is practically the same.

curve, located *quite close* to the eutectic between the microcline components and quartz.

As microcline (from dikes of granite-pegmatite) always contains considerable Ab_{+An}, most often 25–30 Ab_{+An}:75–70 Or, it almost certainly has a somewhat lower melting-point than pure Or, and we must therefore assume that the eutectic Qu:Or contains a little more Qu than the eutectic Qu:microcline.

As the binary eutectic we shall assume 28 Qu:72 Or. As pure Ab has nearly the same melting-point as pure Or, nearly the same molecular weight (Or = 279.4, Ab = 263.3), and possibly also almost exactly the same melting-heat, it must be supposed that the eutectic Qu:Ab holds about the same percentage of Qu as the eutectic Qu:Or. As an approximation we may consequently assume 28 Qu:72 Ab as the binary eutectic.

QU:AN AND QU:AB, QU:OR

For the pressure of one atmosphere we have the synthetic determination: E_{Qu-An} = 48 Qu[1]:52 An, at 1353°. Further we refer to the determinations just mentioned (for a very high pressure):

E_{Qu-Or} = *ca.* 28 Qu:72 Or
E_{Qu-Ab} = *ca.* 28 Cu:72 Ab.

Even if an error of a few per cent may be found in the latter statements, it is evident in any case that the binary eutectic Qu:Or or Qu:Ab contains much less quartz than the binary eutectic Qu:An. This is in accordance with Ab (as well as Or) having an essentially lower melting-point, in round numbers 450°, than An.

The course of the melting-curve (see Fig. 3) on the Qu side in the neighborhood of Qu will be about the same, whether the second compound is An or Ab (or Or). If we extend the curve, experimentally determined for E_{Qu-An} on the quartz side, and draw the curve on the feldspar side for E_{Qu-Ab} (respectively E_{Qu-Or}) about parallel with E_{Qu-An}, a binary eutectic E_{Qu-Ab} (respectively E_{Qu-Or}) will appear with composition about 25–30

[1] Qu here signifies christobalite from the melting-point to 1470° and tridymite from 1470° to 1353°.

Qu:75–70 Ab (or Or), just as we have in reality derived from the analyses of the graphic granites.

As the melting-points as well as the binary eutectics (as will be shown in a following chapter) are only very little displaced by pressure, we are justified in drawing a parallel between the eutectic SiO_2:An, determined for low pressure, and the eutectic SiO_2:Ab (or SiO_2:Or), calculated for high pressure. The case is somewhat

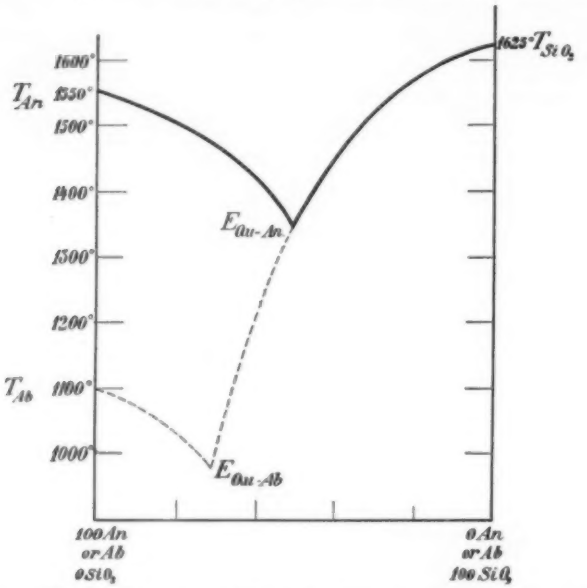

FIG. 3.—Melting-diagram, An:SiO_2 (after Bowen), and Ab:SiO_2 (schematic after Vogt).

complicated, however, by the fact that SiO_2 in one case is tridymite but in the other quartz.

The melting-point for the binary eutectic E_{Qu-Ab} (or E_{Qu-Or}) must, according to the nature of the case, lie considerably lower than the melting-point of pure Ab (or Or), consequently considerably lower than 1100° and certainly somewhat lower than 1000°. As an estimate we set it at 975°, which should be approximately correct.

Qu:Ab+An.—In a *ternary* system consisting of two components (as, for instance, Ab and An), which form a binary mix-crystal system of type I, and a third component independent of the former (as Qu or $CaMgSi_2O_6$), there appear, according to F. A. H. Schreinemaker's theoretical investigations,[1] two melting-surfaces (Fig. 4),

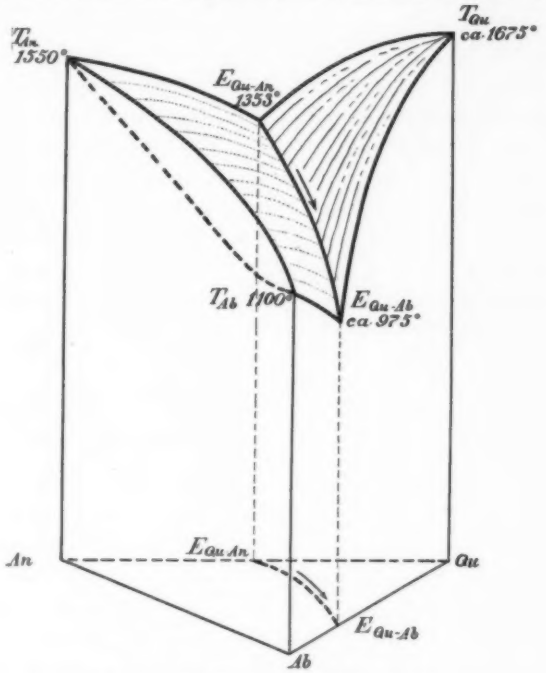

FIG. 4.—The ternary system An:Ab:Qu

which intersect in a curve, viz., "eutectic boundary-line"[2] or briefly a "eutectic line or curve." Three subcases may occur, accordingly as the eutectic boundary-line (E_{Qu-An} to E_{Qu-Ab} on Fig. 4) has a continuous decline, a minimum, or a maximum. In the ternary system Diops:Ab:An the eutectic line according to Bowen's experimental investigations (see Fig. 6) has a continuous decline from $E_{Diops-An}$ (at 1270°) to $E_{Diops-Ab}$ (at a little below

[1] *Zeitschr. f. physikalische Chemie*, 1905, Vols. 50, 51, and 52.

[2] This term I have used in my earlier treatise. Boeke (*loc. cit.*) uses the shorter term "eutectic line."

1100°), consequently with a difference of *ca.* 200° between the two points.

For the analogous system Qu:Ab:An, where the difference between the two points (E_{Qu-An} at 1350° and E_{Qu-Ab} at probably a little below 1000°) is still greater, certainly at least 350°, we may also suppose a continuous decline for the eutectic line. This line, on account of the steep decline near Ab of the binary liquidus curve between An and Ab, will probably assume the course outlined on the horizontal projection, Figure 5. The crystallization between Ab+An and Qu has consequently the same course as between Ab+An and diopside (see below). Even if the curve between E_{Qu-An} and E_{Qu-Ab}, contrary to our conjecture, should show a maximum in the vicinity of E_{Qu-An} or a minimum in the vicinity of E_{Qu-Ab}, this would in no degree worth mentioning modify the course of the curve in the horizontal projection.

We calculate the chemical composition of the end members and of a pair of intermediate compositions.

TABLE III

Percentage of		E_{Qu-Ab}	$E_{Qu-Ab+An}$ by		E_{Qu-An}
			¾ Ab:¼ An	¼ An:¾ Ab	
Ab+An	Qu	28	36	42	48
	Ab	72	48	29	.....
	An	.....	16	29	52
	SiO_2	77.75	75.95	74.5	70.5
	Al_2O_3	13.95	15.2	16.25	19.0
	CaO	.....	3.2	5.85	10.5
	Na_2O	8.50	5.65	3.4	.....

That the calculation here given of the eutectic between quartz and albite, oligoclase, etc. (which is supported by the theoretical argument on the eutectic Qu:An at the pressure of one atmosphere, and by analogy conclusions according to the composition of microcline graphic granite) is essentially correct, is confirmed by the close conformity between the two analyses, Nos. 12 and 13, of oligoclase graphic granite and the compositions here calculated, especially for ¾ Ab:¼ An.

For the system *Quartz*:*Orthoclase* (microcline, with about 72 Or+28 Ab_{+An}):*Albite* (with about 88 Ab or Ab_{+An} and 12 Or) we must have three individualization-fields, with partial eutectics

respectively about 26 Qu:74 orthoclase (72 Or+28 Ab), about 26 Qu:74 Albite (Ab with little An and about 12 Or) and about 40 Or:60 Ab_{+An} (or 42:58). If we leave the inconsiderable

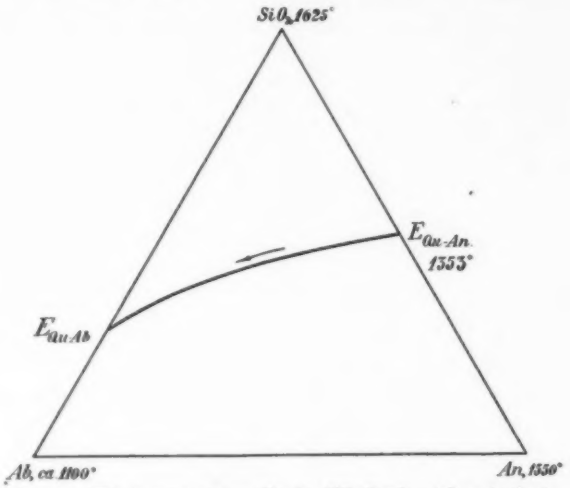

FIG. 5.—The ternary system Ab:An:SiO_2 (horizontal projection)

admixture of An out of consideration, we must suppose for Qu:Or:Ab a ternary eutectic, with about 26 Qu and about 42 Or:58 Ab, consequently about 26 Qu:31 Or:43 Ab (or Ab_{+An}).

We calculate the composition of these eutectics (see *Tscherm. Mitt.*, Vol. XV [1906], p. 385) as shown in Table IV.

TABLE IV

APPROXIMATE CALCULATION OF EUTECTICS QUARTZ: ORTHOCLASE (MICROCLINE): ALBITE (WITH LITTLE AN)

Percentage of	Eutectic Qu:Orthoclase (72 Or:28 Ab+An)			"Ternary Eutectic" Qu:Orthoclase:Albite (with 42 Or:58 Ab+An)				Eutectic Qu:Albite (88 Ab or Ab+An:12 Or)			
Qu	26.0	26.0	26.0	26.0	26.0	26.0	26.0	26.0	26.0	28.0	32.0
Or	53.3	53.3	53.3	31.1	31.1	31.1	31.1	8.9	8.9	8.7	8.6
Ab	20.2	19.2	17.7	42.4	41.4	39.9	36.9	64.6	63.6	60.3	53.4
An	0.5	1.5	3	0.5	1.5	3.0	6.0	0.5	1.5	3.0	6.0
SiO_2	74.7	74.45	74.05	75.55	75.3	74.9	74.15	76.45	76.2	76.45	76.9
Al_2O_3	13.85	14.0	14.25	14.1	14.25	14.55	15.05	14.35	14.5	14.4	14.15
K_2O	9.0	9.0	9.0	5.25	5.25	5.25	5.25	1.5	1.5	1.45	1.45
Na_2O	2.35	2.25	2.1	5.0	4.9	4.7	4.35	7.6	7.5	7.1	6.3
CaO	0.1	0.3	0.6	1.0	0.3	0.6	1.2	0.1	0.3	0.8	1.2

TABLE V

Percentage of	Andesites				Dacite	Spherulit Rock	Dacite	Quartz-porphyry	
	Emmons		(Idd.)	(Lag.)	(Idd.)	(Lag.)	(Lag.)	(Lasp.)	(Streng)
	14a	15a	16a	17a	18a	19a	20a	21a	22a
SiO_2	59.06	60.39	62.00*	62.54	69.36	76.48	75.07	72.24	74.11
Al_2O_3	16.40	16.96	17.84		16.23	12.09	12.15	13.63	13.69
Fe_2O_3	2.88	1.50	.....	23.56	0.88	0.95	1.67	.....	.....
FeO	4.18	3.42	4.40		1.53	.....	.....	3.05†	1.75
MgO	2.63	3.81	2.64	1.15	1.34	0.39	0.14	0.66	0.05
CaO	4.32	5.41	5.37	4.75	3.17	0.64	0.86	0.95	1.38
Na_2O	5.29	3.37	4.29	3.16	4.06	4.89	4.12	2.95	1.54
K_2O	1.49	2.01	1.47	2.43	3.02	3.78	4.57	5.24	5.67
H_2O, ign.	2.06	2.03	1.66	1.75	0.45	0.77	1.34	1.26	0.56
Total	98.31	99.11	100.13	99.35	100.04	99.96	99.92	100.11	99.90
	Phenocrysts								
	(14b)	(15b)	16b	17b	18b	19b	20b		22b
SiO_2	56.25	55.92	56.41	55.42	(65.77†)	62.14	65.49	.....	61.80
Al_2O_3	28.56	28.58	27.39	28.01	21.51	22.30	18.74	.....	19.28
Fe_2O_3	0.77	1.00	0.69‡	1.09	tr.	tr.	tr.	.....	2.02
MgO	1.06	0.55	0.09	tr.	0.00	tr.	.....	.....	0.01
CaO	6.54	6.65	9.87	9.12	5.72	3.29	0.39	.....	.....
Na_2O	5.42	5.66	5.43	5.10	5.92	10.58	3.53	.....	0.68
K_2O	0.61	0.66	0.30	0.79	0.83	1.69	9.45	.....	12.18
H_2O, ign.	0.33	0.69	.....	0.52	0.34	0.20	1.30	.....	0.25
Total	99.54	99.71	100.24	100.05	100.09	100.20	98.90	.....	100.10
	Glass Basis						Groundmass		
	14c	15c	16c	17c	18c	19c	20c	21c	22c
SiO_2	68.11	68.60	69.94	70.19	76.75	74.59	74.96	74.41	74.44
Al_2O_3	15.56	17.27	15.63		12.32	12.88	13.67	13.39	13.51
Fe_2O_3	0.96	2.09	.....	17.19	.....	0.80	1.80	.....	.....
FeO	2.12	0.58	1.89		1.36	.....	.....	3.08	2.25
MgO	1.10	0.40	0.28	0.53	0.00	0.30	tr.	0.50	0.01
CaO	2.91	1.72	2.49	2.50	1.18	0.76	0.62	1.38	1.19
Na_2O	3.43	3.30	3.83	3.30	3.55	3.30	2.70	3.27	1.40
K_2O	2.61	4.61	2.85	3.89	3.98	5.35	4.14	4.18	5.31
H_2O, ign.	2.82	2.08	3.25	2.31	0.54	1.03	1.52	1.04	1.34
Total	99.62	100.65	100.16	99.91	99.68	99.01	99.41	101.55	99.45

* +0.17 TiO_2, 0.29 P_2O_5 in the rock.

† In the rock 0.13 and in the groundmass 0.30 per cent MnO.

‡ is FeO.

EXPLANATION

Nos. 14 and 15: H. Emmons, "Island of Capraja, at Elba," *Quart. Jour. London*, 1893. The two feldspars 14*b* and 15*b* isolated by density 2.67—No. 16: A. Hague and J. P. Iddings, "Volcanoes of Northern California, Oregon, and Washington Territory," *Amer. Jour. Sc.*, 3, Vol. XXVI (1883). Hypersthene-andesite.—No. 17: A. Lagorio, "Über die Natur der Glasbasis sowie der Krystallisationsvorgänge in eruptiven Magma," *Tscherm. min. u. petrogr. Mitt*, Vol. VIII (1887); from Hliniker Valley, Hungary.—No. 18: Hague and Iddings (*loc. cit.*); California. The plagioclase is andesine-oligoclase. The determination of SiO_2 in the plagioclase is too high, owing to impurity.—No. 19: Lagorio (*loc cit.*). Spherulitic rock; from Alausi, Ecuador.—No. 20: Lagorio (*loc. cit.*); from Summit County, Colorado. With phenocrysts of quartz and two feldspars, one monocline (analysis No. 20*b*) and the other tricline.—No. 21: Laspeyres (see Zirkel's *Textbook of Petrography*, 1894, Vol. II, p. 177); from Halle, Germany.—No. 22: A. Streng, *Neuer Jahrb. f. Min., Geol. u. Pal.*, 1860, from the Harz Mountains.

In order to establish that the crystallization in the eruptive magmas of the different feldspars and of quartz is in conformity with the physicochemical details which we here have developed essentially on the basis of the analysis of graphic granite, we refer *inter alia* to my earlier statement in *Tscherm. Mitt.*, Vol. XXV (1906).

On pages 340 and 342 we give a small selection (analyses Nos. 14–29) from the numerous analyses, compiled from the literature, partly of porphyritic rocks, with special analyses of (*a*) the whole rock, (*b*) the porphyritic feldspar, and (*c*) the glass or groundmass, and partly of some granites with special analyses of (*b*) the basic concretions (or orbicules), and (*c*) the inclosing rock.

Granites with basic concretions (Nos. 24*b*–26*b*), or basic orbs (Nos. 27*b*–29*b*) are shown in Table VI on p. 342.

We call special attention to the following:

1. In the intermediate and the acid eruptive rocks, which contain the ordinary admixture of ferromagnesian silicates and iron, or titanic iron, ore (especially magnetite and ilmenite), an essential part of these minerals crystallizes at an early stage. A small quantity of Fe_2O_3, FeO, and MgO, however, is left in the remaining magma. This appears in the solidified rocks as the glass basis or groundmass in the porphyritic rocks, or as the intervening mass between basic concretions or orbicules in granites.

In the final, very complicated eutectic, chiefly consisting of feldspars and quartz, there usually seem to appear 0.5 per cent Fe_2O_3, 0.25 per cent FeO, and 0.1–0.25 per cent MgO; the figures,

TABLE VI

Percentage of	Basic Concretions					Basic Orbs			
	23*b*	24*b*$_1$	25*b*$_1$	25*b*$_2$	26*b*	27*b*	28*b*	29*b*$_1$	29*b*$_2$
SiO_2	53.80	54.73	56.01	56.53	64.39	55.72	65.57	61.10	68.02
TiO_2	0.77	tr.	1.13	1.40	tr.	0.57	.....	0.51	.....
Al_2O_3	19.20	14.02	15.19	16.47	15.99	21.35	17.46	15.55	15.31
Fe_2O_3	7.60	2.34	2.34	1.58	1.47	.....	4.15	2.10	0.59
FeO		4.92	4.89	5.40	5.98	8.81		2.21	2.14
MnO	.....	.....	0.40	0.20	tr.	0.36	.....	.....	.....
MgO	4.80	7.40	4.67	2.67	1.67	0.63	2.53	6.30	3.41
CaO	5.70	10.20	4.85	4.00	2.57	5.10	2.49	1.12	1.53
Na_2O	2.16	2.98	5.66	5.59	4.96	5.71	2.14	1.23	2.85
K_2O	5.08	2.67	2.16	3.80	2.46	1.23	4.23	9.38	5.67
P_2O_5	1.20	tr.	0.53	0.27	tr.	.....	.....	.....	.....
Ign.	1.28	1.23	1.26	0.93	0.95	0.46	1.26	1.80	0.46
Total	101.59	100.99	99.21	99.98	100.44	99.94	99.83	100.58	99.98

	Intervening Mass								
	23*c*	24*c*	25*c*$_1$	25*c*$_2$	26*c*	27*c*	28*c*	29*c*	
SiO_2	74.40	70.44	71.90	73.69	73.70	70.05	68.27	71.71	.....
TiO_2	.....	.....	0.35	0.28	tr.	0.19	.....	.....	.....
Al_2O_3	13.91	15.63	14.12	12.46	14.44	14.78	15.59	15.05	.....
Fe_2O_3	1.39	1.34	1.20	1.21	0.43	.....	2.13	1.11	.....
FeO		1.12	0.86	1.75	1.49	3.37		0.29	.....
MnO	.....	.....	0.05	0.15	tr.	0.22	.....	.....	.....
MgO	0.28	0.55	0.33	0.17	tr.	0.44	1.19	0.56	.....
CaO	0.61	1.98	1.13	0.36	1.08	3.42	1.93	1.42	.....
Na_2O	4.65	4.03	4.52	4.47	4.21	3.10	3.21	3.39	.....
K_2O	4.36	5.18	4.81	4.92	4.43	4.13	5.37	5.43	.....
P_2O_5	.....	.....	0.11	0.04	tr.	.....	.....	.....	.....
Ign.	0.65	0.55	0.60	0.38	0.61	0.42	1.56	0.61	.....
Total	100.25	100.82	100.35	100.09	100.39	100.12	99.25	99.77	.....

EXPLANATION

No. 23: See Rosenbusch, *Elemente der Gesteinslehre*, 1901; from Pelvoux.—No. 24: Graber, *Jahrb. d. k.-k. geol. Reichsanstalt*, Wien, 1897; from Topla in Carinthia.—No. 25: Clarke, *U.S. Geol. Survey Bull. 168*, 24; from Mount Ascutney, Vt.—No. 25b_2 and c_2, granite porphyry.—No. 26: J. A. Phillips, *Quart. Jour. London*, 1880; from Peterhead, Scotland.—No. 27: H. Bäckström, *Geol. Fören. Förh.*, Stockholm, 1894; from Kortfors, Sweden.—Nos. 28–29: K. v. Chrustschoff, "Über holokrystalline makrovariolithische Gesteine," *Mém. de l'académie des sc. de St. Pétersbourg*, 1894. No. 28 from Altai; No. 29 from Fonni, Sardinia.

however, especially for MgO, depend somewhat upon the Mg-bearing mineral component in question (biotite, hypersthene, etc.).

2. As previously mentioned and as illustrated by analyses Nos. 14–19 and 20–22, plagioclase crystallizes when there is a surplus of Ab+An; orthoclase, on the contrary, when there is a surplus of Or in the original solution. The boundary lies, as previously explained, at about 0.4 Or:0.6 Ab+An. In the plagioclase which crystallized first, relatively much An appears, consequently relatively much CaO (and Al_2O_3) (cf. the analyses Nos. 14*b*–19*b*). In consequence, the remaining magma shows a decreasing percentage of CaO, and this in the "granite eutectic" falls to 0.25, 0.5, or 1 per cent CaO, or, with predominating plagioclase in the eutectic, not quite so low.

If we leave magmas with only a trifle of Na_2O (Ab), or of K_2O (Or), out of consideration—where the crystallizing orthoclase absorbs practically all of the Ab or the crystallizing albite or albite-oligoclase the Or—the contents of Or (or K_2O) in the magma remnant increases by the crystallization of plagioclase (cf. the analyses Nos. 14*c*–19*c* and 24*c*, 27*c*), and the contents of Ab (or Na_2O) in the magma remnant increases by the crystallization of orthoclase (cf. No. 21*c*). But we especially emphasize that this *relative increase has a limited course*, and that *the limit of about 0.4 Or:0.6 Ab or Ab+An is not exceeded*, or only perhaps now and then somewhat exceeded because of supersaturation. In this matter we especially refer to the analyses of the glass basis or groundmass of porphyritic rocks. In judging these analyses, however, we must take into consideration that the glass, as is shown, for example, by the water content, is always or nearly always somewhat decomposed, whereby especially a little alkali will be extracted. We further refer to the intervening material between basic concretions or orbicules in granites.

As examples these intervening masses from casually chosen granites show:

In Nos. 28 and 29, where orthoclase crystallized first, respectively 0.45 Or:0.55 Ab_{+An} and 0.46 Or:0.54 Ab_{+An}.

In No. 23, where both orthoclase and plagioclase seem to have crystallized at an early stage: 0.37 Or:0.63 Ab_{+An}.

In Nos. 24, 25c^1 and c^2, 26, and 27, where plagioclase (or perhaps predominant plagioclase and some orthoclase) crystallized at an early stage, respectively: 0.40, 0.38, 0.41, 0.38, and 0.36 Or. The remainder is Ab_{+An}.

These values, calculated from the analyses, for the proportions $Or:Ab_{+An}$ need, however, a small correction, as we have taken for granted that the whole quantity K_2O, Na_2O, and CaO form respectively Or, Ab, and An, while in reality a trifle K_2O (and a still smaller amount of Na_2O) in several cases enters into biotite.

3. The glass basis, or the groundmass, in andesites (with at least about 56 per cent SiO_2, that is to say, with at least so much SiO_2 that a little of the independent component quartz entered into the melted magma) and also in dacites, trachytes, rhyolites, quartz-porphyries, etc. (with max. about 72 per cent SiO_2) without exception shows an increased percentage of SiO_2 as compared with the entire rock. Gradually as the crystallization, for example, of an andesite with 59 per cent SiO_2 (No. 14*a*) advances with solidification of magnetite, ferromagnesian silicates, and medium-basic plagioclase, the temperature drops, and simultaneously the mother liquor becomes richer in SiO_2. By rapid cooling the viscosity increases and causes the cessation of the crystallization; in other words, the fluid remnant stiffens into glass, with varying percentages of SiO_2 according to the time at which the crystallization ceased. And this point of time may lie even considerably lower than the stage of the final eutectic. This we may illustrate by giving the percentages of SiO_2 in the entire rock and in the glass basis (or in some cases the groundmasses); the latter, however, always contains a little H_2O, showing that it is somewhat decomposed. (See my treatise in *Tscherm. Mitt.*, Vol. XXIV [1905].)

ANDESITES

Percentage of SiO_2 in									
Rock	56.8	57.8	58.1	59.1	60.1	60.4	62.0	62.3	62.5
Glass basis	64.5	65.1	70.8	68.1	68.7	68.6	69.9	67.0	70.2

DACITES

Percentage of SiO_2 in						
Rock	65.6	68.3	69.4	65.5	65.3	67.3
Glass basis	71.9	74.8*	76.75	70.2	70.2 and 73.6	72.4

*Goundmass.

Further we include a series of rhyolites, dacites, spherulites, and quartz-porphyries, in which phenocrysts of feldspar (orthoclase or plagioclase) as well as of quartz usually appear. Gl. = glass basis; Grm. = groundmass; Sph. = Spherulite.

TABLE VII

	Percentage of	Dacites		Rhyolites				Spherulitic Rocks	
		Sph.	Grm.	Grm.	Sph. and Gl.	Gl.	Sph. Gl.	Gl.	Sph. Gl.
Rock	SiO_2	74.6	75.07	72.5	75.8	73.0	71.4	76.5	72.8
Gl., Grm., or Sph...	SiO_2	76.05*	74.96	77.5†	72.7	72.6	74.5	72.5	74.6	73.7	70.7
	H_2O	1.2	1.5	1.4	0.9	4.55	1.4	4.4	1.0	1.4	3.7

*Only 10.24 per cent Al_2O_3. Somewhat decomposed.
†Only 11.52 per cent Al_2O_3. Somewhat decomposed.

	Percentage of	Spherulitic Rocks with Spherulites and Glass Basis					Quartz-Porphyries with Groundmass			
Sph.....	SiO_2	73.2	74.4	75.4	73.4	Qu-porph...	72.0	72.2	74.1	75.2
Gl.....	SiO_2	72.4	72.7	73.0	73.1	Grm.....	74.0	74.4	74.4	(76.8?)
	H_2O	1.4	1.1	3.6	0.9			1.0	1.3	0.8

The glass basis, or the groundmass, in the porphyritic rocks consequently shows:

a) In rocks with about 60 per cent SiO_2, a sometimes very considerable increase in the percentage of SiO_2 (for example, from 58.1 to 70.8 per cent SiO_2) and in rocks with about 65 or 65–70 per cent SiO_2 a smaller increase, though in undecomposed state not above 75 per cent SiO_2.

b) In rocks with about 73–75 per cent SiO_2 we find, on the other hand, about the same percentage of SiO_2 in the glass basis or groundmass as in the entire rock. The analyses show some slight variations, partly in one and partly in the other direction. But this is certainly caused in some cases only by slight inaccuracies in the relatively old analyses, and in others by the groundmass and especially the glass basis (probably without exception) being a little decomposed, as shown by a little H_2O.

These law-governed relations, which are established by numerous analyses, may depend on the fact that in a magma consisting chiefly of Qu and Or, Ab and An components, with a

surplus of feldspar components, the crystallization of feldspar may continue without a simultaneous secretion of quartz until the eutectic boundary-line between quartz and the feldspar components has been reached. When this has occurred, however, a simultaneous crystallization of feldspar and quartz commences, with only a quite inconsiderable change of the SiO_2 percentage of the magma remnant, while we constantly more and more approach the "ternary" eutectic: Qu:Or:Ab+An (with a trifle magnetite and ferromagnesian silicate).

The groundmass in the quartz porphyries and the closely related rocks consists, as is well known, in some cases of microfelsite and in others of granophyre, and these structural forms indicate a *simultaneous* crystallization of quartz and the feldspar in question. The final crystallization consequently took place also with regard to the structure at a eutectic or eutectic boundary-line.

Especially acid quartz porphyries (with more than 75 per cent SiO_2) show phenocrysts of quartz and feldspar in about equal amounts, but groundmasses of normal microfelsite or granophyre, that is to say, with relatively less quartz than among the phenocrysts. The groundmass, consequently, here must have grown a little more basic than the original rock. I lack material, however, to prove this by chemical analysis.

If we now turn to the deep-seated igneous rocks, we find that the quartz-norites and quartz-gabbros (with about 1 to 5 or 6 per cent quartz), the quartz-diorites, the quartz-syenites, etc., prevailingly show that the quartz first began crystallizing at a relatively late stage. As we shall explain later (Figs. 17 and 18) when treating of the quartz-norites, this crystallization of quartz at a late stage took place, not by itself, but simultaneously with the final crystallization of the feldspar (the plagioclase) and the ferromagnesian silicate in question. In the granite porphyries, which contain but little ferromagnesian silicate and magnetite but are especially rich in feldspar, with about 66–70 per cent SiO_2, the crystallization commenced with the solidification of some feldspar.

The crystallization in ordinary granites usually commenced with a solidification of some magnetite and ferromagnesian silicate,

while the feldspar first commenced crystallizing at a somewhat later stage. The sequence of the *commencement* of the crystallization in the granites is in most cases (1) pyrite, zirkon, apatite, etc.; (2) iron ore and ferromagnesian silicate; (3) feldspar; (4) quartz. But it appears from the structure that the ferromagnesian silicate, especially biotite, continued crystallizing after the commencement of the solidification of both feldspar and quartz, and that the feldspar continued crystallizing also during the segregation of the quartz. In most of the granites, however, we are unable to determine with accuracy from the structure, the quantitative proportions of feldspar and quartz during the intermediate and later stages of the crystallization.

The case is complicated by the fact that the granite magma contains, besides the usual ferromagnesian silicate, Or, Ab, An, and Qu components, some H_2O, probably partly entering into a SiO_2 combination, for example, as H_2SiO_3 (?), and the latter was not split up until a later stage of the crystallization period. If this supposition is correct, the consequence will be a somewhat reduced quantity of the independent quartz component during the first part of the crystallization period—that is to say, during the first part of the crystallization the feldspar was relatively more abundant than that corresponding to the proportion calculated from the relation between feldspar and quartz in the resulting solid rock.

We have an instructive orientation on the composition of granite magmas at a late stage of the solidification in the composition of the intervening masses between basic concretions, or orbicules, in granites, which show these structural elements. (See analyses Nos. 23*c*–29*c*.) These intervening masses prove throughout that during the crystallization a displacement of the composition of the magma remainder took place in the direction of the—in other ways determined—"granite eutectic," and we especially emphasize that the analyses of the intervening masses Nos. 23*c*, 25*c*$_2$, and 26*c* almost exactly correspond with the "granitic eutectic."

Above we have only considered granites with relatively basic concretions, or orbicules. But from strongly *acid* granites, with about 78–80 per cent SiO_2 in the whole rock, we know a couple of

examples[1] of orbicular structure, the orbs chiefly consisting of *quartz*, and in addition some sillimanite and tourmaline (!). The intervening mass between the orbs here contains less SiO_2 (and less quartz) than the orbs, therefore reckoned from the acid pole there here appears a displacement of the residual magma in the direction of the eutectic, quartz:feldspar. In these cases we have, however, the complication that the orbs contain much tourmaline and sillimanite, the latter being very little soluble in acid magma.

Finally we compile a series of analyses of glass bases (Gl.), respectively groundmasses (Grm.), and spherulites (Sph.) from porphyritic rocks, and intervening masses from granites with basic concretions, etc. These analyses represent approximately the composition of the residual magmas resulting from far-advanced

TABLE VIII

ANALYSES OF THE FINAL PRODUCT OF THE CRYSTALLIZATION OF ACID ROCKS—THE GRANITIC EUTECTIC

	No.	SiO_2 Without H_2O	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	Total
Gl., Sph.	30	73.4	72.69	.....	15.04	....	tr.	0.25	1.75	8.85	0.94	**99.52**
Grm.	31	74.0	74.0	.....	.....	.....	.....	.....	0.60	7.57	.....	.....
	32	73.4	72.44	.....	16.28	....	0.20	0.59	2.12	6.49	1.35	**99.47**
	33	74.4	73.57	13.80	.....	1.54	0.26	0.99	3.09	5.74	1.08	**100.07**
Glass basis.	34	73.3	72.35	13.97	1.29	....	0.46	0.72	3.58	5.38	1.37	**99.12**
	35	73.7	73.05	14.67	0.89	....	0.26	0.97	3.99	5.11	0.91	**99.85**
	19c	75.3	74.59	12.88	0.80	....	0.30	0.76	3.30	5.35	1.03	**99.01**
	25c$_a$	74.0	73.69	12.46	1.21	1.75	0.17	0.36	4.47	4.92	0.38	**100.09**
Intervening mass of granites	23c	74.9	74.40	13.91	1.39	....	0.28	0.61	4.65	4.36	0.65	**100.25**
	26c	74.1	73.70	14.40	0.43	1.49	tr.	1.08	4.21	4.43	0.61	**100.39**
	36	74.0	73.21	12.90	2.10	....	0.27	0.88	4.83	4.75	1.04	**99.98**
Spherulites	37c$_1$	75.1	74.52	12.97	2.02	....	0.25	0.92	4.26	4.53	0.83	**100.30**
	38	74.7	73.72	12.91	1.37	....	0.25	1.37	4.02	4.45	1.36	**99.45**
Gl.	39	73.5	72.70	13.79	1.01	....	0.65	2.07	4.93	4.33	1.10	**100.48**
Sph.	37c$_2$	74.0	73.42	14.29	1.01	....	0.43	1.00	5.61	3.19	0.84	**99.79**
	40	74.8	74.36	14.46	1.62	....	0.44	1.49	6.11	1.49	0.57	**100.54**

[1] From Kragerö and Modum in Norway and from Pine Lake in Ontario (lecture by W. C. Brögger on the Kragerö locality, in *Kristiania Videnskr. Selsk*, 1901, and Frank D. Adams, *Bull. Geol. Soc. Amer. No. 9*, 1898, see résumé in my treatise in *Tscherm. Mitt.*, Vol. XXV [1906]).

EXPLANATION

No. 30: Spherulitic glass basis from liparites.—No. 31: groundmass from quartz-porphyry.—Nos. 32, 33: Glass basis from trachytes.—Nos. 34, 35, 39: Glass basis from spherulitic rocks—Nos. 36, 37c_1 and c_2, 38, and 40: Spherulites from spherulitic rocks.

Nos. 30, 32, 34, 35, 36–40 reprinted from the above-cited excellent treatise of Lagorio, 1887.—No. 31, see Zirkel's textbook.—No. 33, Williams, *Neues Jahrb. f. Min., Geol. u. Pal.*, Beil., Bd. V, 1887 (see also my treatise in *Tscherm. Mitt.*, Vol. XXIV [1905]).

crystallization. In judging these analyses we must take into consideration that throughout a little alkali was probably extracted from the glass basis, so that the determined percentages of alkalies may be a trifle too low. The analyses are generally arranged according to decreasing K_2O (Or) or increasing Na_2O (or Ab+An).

We especially direct attention to the close accordance between these analyses of the residual magmas resulting from the solidification—partly from dike and surface rocks, and partly from deep-seated rocks[1]—and the eutectic Qu:Or:Ab+An, calculated on the basis of the graphic-granite analyses and the theoretical explanations. (See the table, p. 339.) The accordance is especially pronounced when we take into consideration that the percentage of SiO_2 in the analysis of graphic granite will be reduced about 1 per cent when 1 or 2 per cent of ferromagnesian silicate and magnetite is added, and that the analyses of groundmasses, etc., which are rich in plagioclase, contain a little more CaO (or An) than the graphic granites calculated in the table, page 339, where only a small admixture of An is presupposed. The analyses in the table, page 348, of the residual magma represent the *granitic eutectic*, consisting of predominant Qu, Or, and Ab+An, with the addition of quite small admixtures of iron oxide (Fe_3O_4) and ferromagnesian silicate.

If we leave the latter quite subordinate admixtures out of consideration, the analyses Nos. 25c_2, 23c, 26c, 36, 37c, and 38 almost exactly represent the "ternary" eutectic Qu:Or:Ab (or Ab+An) with nearly exactly 0.4 Or:0.6 Ab (or Ab+An). And this we may by a short catchword name the "*ternary*" *granitic*

[1] Regarding the inconsiderable influence of the pressure on the composition of the eutectic we refer to a following chapter.

eutectic. This term strikes the essential point, since the eutectic in question consists practically only of Qu, Or, and Ab. To this must, however, be added a quite small admixture of An, iron oxide, and ferromagnesian silicate, so that the eutectic in reality is more complicated. In order to avoid misunderstanding I have therefore put the term "ternary" in quotation marks.

[*To be continued*]

RUSSELL FORK FAULT OF SOUTHWEST VIRGINIA[1]

CHESTER K. WENTWORTH
University of Iowa

INTRODUCTION

The nature of the northeastern termination of the great overthrust block of the earth's crust, bounded on three sides by the Pine Mountain fault, the Hunter Valley fault, and the Jacksboro cross fault of Tennessee, has long been an unsolved problem to students of Appalachian structural geology. Many geologists have noted the rather abrupt ending, near the breaks of Big Sandy River, of the imposing barrier of Pine Mountain and its replacement to the northeastward by the irregular ridges and valleys of the unbroken coal field, but the manner in which the great anticlinal fold and the resulting thrust fault died out has not until recently been satisfactorily solved.

In 1916 Hinds,[2] in his report on the Clintwood and Bucu quadrangles, called attention to a zone of disturbed rocks nearly at right angles to the general lines of disturbance in this region and extending partly across the trough of coal-measure rocks from Big A Mountain to Skegg Gap on Pine Mountain. Hinds attributed the disturbance in this zone to the same forces that produced the Hunter Valley fault on the southeast and the Pine Mountain fault on the northwest, but he failed to perceive its significance, for he thought it was limited to certain areas and did not extend entirely across the synclinal block.

In April, 1920, Mr. M. R. Campbell, in charge of geologic work in this coal field for the United States Geological Survey, called attention to the possibility of the belt of disturbed rocks mapped

[1] Published by permission of the Directors of the U.S. Geological Survey and the Virginia Geological Survey. The illustrations were prepared for the Virginia Geological Survey.

[2] Henry Hinds, "The Coal Resources of the Clintwood and Bucu Quadrangles, Virginia," *Virginia Geol. Survey Bull. 12* (1916).

by Hinds being but part of a continuous fault or zone of faulting from the Hunter Valley fault at Big A Mountain to Skegg Gap in Pine Mountain, and the author was requested to examine the region as carefully as the limited time at his disposal would permit, in order to establish the character and extent of the movements that produced the disturbance. The result of his examination was the establishment of the presence of an overthrust fault entirely across the great crustal block, thus showing that it is bounded on all four sides by overthrust faults and that it has moved bodily to the northwest a distance of many miles. The results of his studies and their application to the mechanics of the problem of the overthrusting of this great mass of strata for at least six miles are here set forth.

The fault bounding the crustal block on the northeast, which, on account of its general agreement with the course of Russell Fork, is here called the Russell Fork fault, was mapped in connection with coal investigations carried on co-operatively by the Virginia Geological Survey and the federal Geological Survey. The areas mapped as undifferentiated buckled and faulted rocks by Hinds,[1] on the Clintwood and Bucu quadrangles, were subjected to careful study by the writer to determine whether or not there was a continuous break across the coal-measures trough from the vicinity of Big A Mountain to Skegg Gap, but in the two weeks spent on this study there was not time to cover much of the area lying on either side of this zone and the structure contour maps of the report by Hinds furnished many data in compiling the sections shown below and in deducing the amount of displacement.

The writer is indebted to Mr. M. R. Campbell for many helpful suggestions and much assistance in the course of the study.

Hinds,[2] in his report on the coal resources of the Clintwood and Bucu quadrangles, describes the structure of the northeastern end of the Middlesboro syncline in considerable detail. His studies here and farther northeast in Buchanan County[3] have shown that

[1] *Op. cit.*

[2] *Op. cit.*

[3] Henry Hinds, "Geology and Coal Resources of Buchanan County, Virginia," *Virginia Geol. Survey Bull. 18* (1918).

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the great overthrust of Pine Mountain suddenly becomes very much less severe at Skegg Gap, and from there northeastward the structure is essentially a low anticline broken by a minor overthrust which decreases rapidly in extent of thrust and comes to an end a few miles into Buchanan County. He considered that the principal Pine Mountain overthrust was cut off at the northeast end by the Skegg Gap fault which he mapped as far as Russell Fork. Between this point and Big A Mountain he has mapped a number of narrow areas of faulted and buckled rocks which he describes in some detail and in explanation of which he postulates lateral shearing with the southwest side moving northward with some overthrusting and buckling against the northeast side. He states that succeeding this movement there was normal faulting along this line in which the southwest side was downthrown. His evidence for this belief is not clear, and his several areas of disturbed rocks are separated by areas in which he found no evidence of movement.

DESCRIPTION OF THE CUMBERLAND BLOCK[1]

The structure of the area concerned in this paper has been described in considerable detail at many points by previous writers.[2] It is not the writer's purpose to give here a thorough description of the structure or topography but rather to point out briefly their alient features.

The "remarkable quadrilateral block" whose southwestern extremity was first recognized by Safford and described in detail by Keith extends from the valley of Cove Creek in Campbell County, northeastern Tennessee, northeastward for one hundred and twenty-five miles to the valley of Russell Fork of Big Sandy River in Dickinson and Buchanan counties, Virginia. It is surprisingly uniform in width, averaging about twenty-five miles, is bounded on the northwest by the Pine Mountain fault and on the southeast by the Hunter Valley fault and the closely associated Wallen Valley fault, which is developed only from near Big Stone

[1] This name is here applied for the first time.

[2] J. M. Safford, *Geology of Tennessee* (1869); M. R. Campbell, *Geologic Folios 12 and 59*, Arthur Keith, *Geologic Folios 33 and 75*, G. H. Ashley and L. C. Glenn, *Prof. Paper 49*, all of the *U.S. Geol. Survey.*

FIG. 1.—Areal map of the Cumberland block showing its location in the states of Virginia, Kentucky, and Tennessee and locations of the structure sections of Figures 3 and 4.

Gap southwestward. The southwest end is terminated by the Jacksboro cross fault and the northeast end by the Russell Fork cross fault to be described below in more detail. The general relations of these boundary faults may be more clearly seen by reference to the map and diagram, Figures 1 and 2, and to the structure sections, Figures 3 and 4.

From Norton, Virginia, northeastward, coal-measure rocks are exposed at the surface throughout the entire width of the block; but from Norton southwestward the block may be divided into two parts, that part lying northwest of Stone and Cumberland mountains being synclinal in structure and composed of coal-measure rocks, whereas that part lying southeast of these mountains is anticlinal in structure and composed of rocks of very much greater age. The syncline, which is now generally known as the

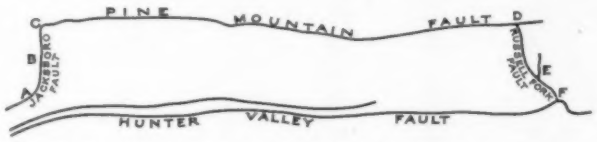

FIG. 2.—Outline diagram of Cumberland block showing the bounding faults

Middlesboro syncline, is a broad, flat-bottomed trough at the northeast end of the block, but farther west in the vicinity of Dante an arch appears which develops rapidly westward into the Powell Valley anticline that constitutes the southern part of the block. Both the Middlesboro syncline and the Powell Valley anticline are characterized by steep dips on the northwest limbs and gentle dips on the southeast limbs, as shown in the sections. The erosion by Powell River and its tributaries of the rising crown of the Powell Valley anticline accounts for the exposures of pre-Carboniferous rocks in the southern portion of the block and the narrowing of the coal-measure portion on the north. The general relation of these different structural units and their expression in the areal relationships of the Pennsylvanian and pre-Pennsylvanian rocks may be seen by reference to Figure 1.

The topography of the block is intimately related to the structure. Pine Mountain throughout its entire length is a conspicuous

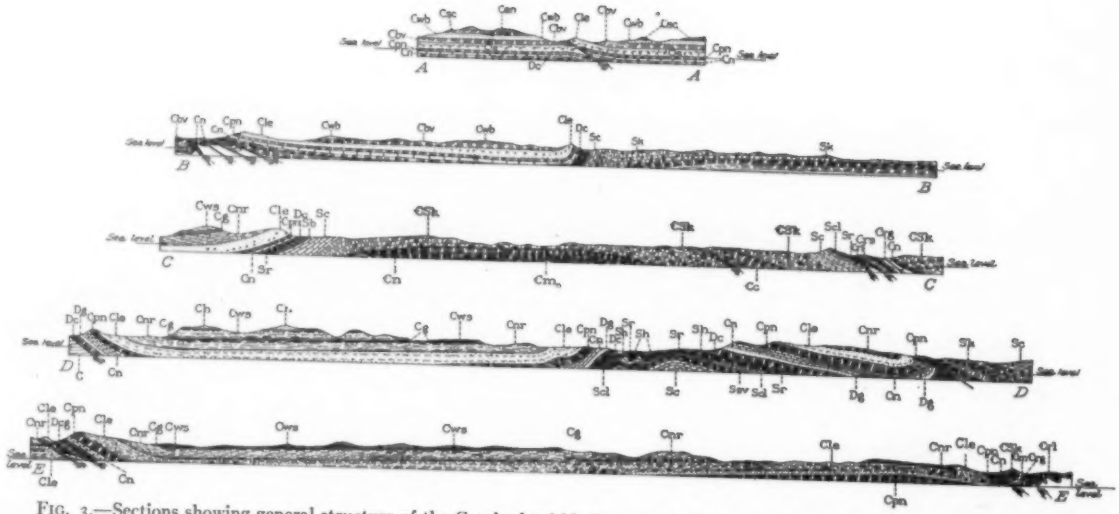

FIG. 3.—Sections showing general structure of the Cumberland block. Sec. *A–A* is seen as looking northwest, the others as looking northeast. For locations see Figure 1. (These sections have been compiled and somewhat generalized from *Geologic Folios 12, 33, 59, and 75, Professional Paper 49,* and *Bulletin 541 f.*, all of the U.S. Geol. Survey.)

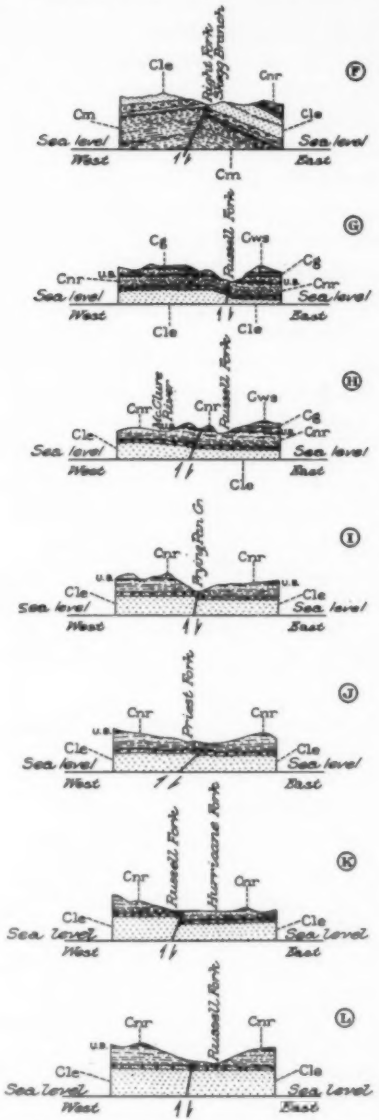

FIG. 4.—Sections showing the character of the Russell Fork fault at several places. For locations see Figure 1. Sections are shown as looking north.

barrier especially on its northwest face, where at many points it rises in less than a mile one thousand to two thousand feet above the streams which parallel it. For nearly ninety miles no stream crosses it and in the entire distance of one hundred and twenty-five miles not over half a dozen roads afford passage from one side to the other. Its crest is the resistant conglomerate of the Lee formation, which marks the edge of the overthrust block. Cumberland Mountain and Stone Mountain are composed of the same formation, which is steeply upturned in that part of the fold common to the syncline and anticline mentioned above. Black Mountain and parts of Sandy Ridge are residual mountains left in the dissection of the nearly horizontal coal measures. Big A Mountain, according to Hinds,[1] is composed of resistant sandstones of the Rockwood formation, overthrust on the coal measures.

The surface features of that portion of the block within the coal field are those of a maturely dissected plateau with sharp-crested ridges and V-shaped valleys, for the most part without valley flats. The surface of the pre-Carboniferous portion is rolling, with sink holes in the limestone portion and some, though not at this particular point very striking, allineation of ridges and valleys with the northeast-southwest trend of the Appalachian structure. The surface configuration of the area and the control of topography by the great structural features is admirably shown on the contour maps of the United States Geological Survey, to which the reader is referred for further detail.

THE RUSSELL FORK FAULT

The Russell Fork fault differs from the faults which bound the Cumberland block on the other three sides in that it is not a low-angle overthrust and that in it the greatest displacement is in a horizontal direction with comparatively little vertical movement. Its trace[2] is closely followed except at a few places by Russell

[1] Henry Hinds, "The Geology and Coal Resources of Buchanan County, Virginia," *Virginia Geol. Survey Bull. 18* (1918), pp. 58–59.

[2] "Trace" of a fault is here used in its mathematical sense of the line of intersection of one surface with another, i.e., the intersection of the fault plane with the surface of the earth.

Fork of Big Sandy River, and even at those places it is marked by the allineation of minor drainage lines or surface features which would not otherwise be easily explained.

Erosion of crushed and weakened rocks along the fault trace has produced the low saddle at Skegg Gap and the saddle in the point of the spur west of Russell Fork and one mile north of B.M. 1221[1] on the Clintwood quadrangle.

From a point one-half mile upstream from B.M. 1282 to B.M. 1221 Russell Fork flows in a course somewhat farther northeastward than in adjacent parts of its course up and down stream. In the high land which lies southwest of this part of the river and northwest of the village of Haysi are cut two short cleftlike hollows which are closely aligned with the fault trace as located to the north and south, and have without doubt been determined by the presence of the weaker rock in the zone of deformation adjacent to the fault. One of these hollows enters the river valley just at the railroad bridge east of B.M. 1221 and the other extends from near Haysi south and just to the west of B.M. 1380. These hollows are somewhat straighter and narrower than most of the ravines of similar size which erosion has cut in the rocks of this region, but their most distinctive characteristic is their location where they cut off in part the narrow strip of high land between them and the river. Taken together and with the other topographic features which show alignment, they are very significant, and their locations are not to be explained as accidental.

Between McClure River and Russell Fork, at the close approach before they join, a low saddle in the spur owes its position to the weakness of the rocks along the fault line. The very straight course of Fryingpan Creek from elevation 1,311 feet to its mouth is determined by the fault, and it is interesting to note that this creek has a very slight fall in this part of its course and its bed is graded for the entire distance with ripple-marked sand. Russell Fork leaves the fault trace at a number of points and because of its cutting across the undisturbed and more resistant rocks at

[1] The area crossed by Russell Fork fault is shown in detail on the Regina, Ky., and the Clintwood and Bucu, Va., sheets of the *Topographic Atlas of the United States*. Frequent reference is made to points on these maps in locating the features described.

these points is not so perfectly graded. But the weakness of the disturbed rocks close to the fault has apparently enabled Fryingpan Creek, though a comparatively small stream, to grade its lower course to the temporary base as determined by the rocks over which Russell Fork flows.

The main line of the fault passes somewhat to the north of Abners Gap; from elevation 1,424 feet southeast to the mouth of Carroll Presley Branch it follows Russell Fork for most of the distance, and thence southeast to the point where it is truncated by the main overthrust fault; in the north face of Big A Mountain its trace lies somewhat to the north of the channel of Russell Fork.

A branch leaves the main fault at elevation 1,424 feet, and extends northwestward along the course of Russell Fork to a point in Little Pawpaw Valley about a mile north of Cannady Post Office, and is here named the Little Pawpaw fault.

Along most of its course the rocks northeast of the fault are horizontal or nearly so and undisturbed. The fault plane, or, better perhaps, the planes of movement, for the most part dip at high angles, 75° to 90° to the southwest. That there has been intense compression is shown by the mashed condition of the shale and jointed condition of sandstone on the southwest side of the fault. At numerous exposures in the zone of faulting, slickensides indicate considerable vertical movement which has resulted in lifting the beds on the southwest above those on the northeast, displacing the coal beds by from 50 to 200 feet. Because of the shearing which has brought anticlines into contact with synclines and vice versa, it is difficult to determine the true amount of differential vertical movement, but the essential point is that the vertical movement is slight and that the hanging wall has moved up as a result of thrust. At many points there are planes other than those which bear the vertical slickensides, a series of horizontal slickensides trending closely in the direction of the fault, and usually these surfaces are rubbed and planed much smoother and more nearly plane than the others, indicating, it seems to the writer, that these surfaces are the result of more extensive movement along the fault line than the other planes along which

a slighter movement has taken place, to accommodate the thrust in a direction lateral to the main Russell Fork fault line (Figs. 5 and 6.)

At a few points, notably at Skegg Gap (Fig. 7), the slickensides and planes of movement within the zone of faulting indicate a combination of the main southeast-to-northwest thrust with the side or southwest-to-northeast thrust, making the direction of movement a resultant of the two. The slickensided surface here shows the result of pronounced movement and the white quartz pebbles which are so numerous in the Lee formation are planed off flush

FIG. 5.—View of river bank looking southeast near B.M. 1221. Massive and undeformed sandstone on left of fault trace with deformed shale at right. Strong horizontal slickensides are formed on the face of the sandstone at this place.

with the matrix. It is important to mention that a short distance northwest of Abners Gap the writer saw the most abundant evidence of thrust in the mashing and crushing of shale in a fine exposure and that here the slickensides indicate movement at an angle of 45° to the horizontal in a due north direction. At this point the fault trace is more nearly athwart the main direction of thrust, and here, if anywhere, would be expected evidence of strong overthrust.

In the Little Pawpaw fault there was seen little indication of shearing but abundant evidence of compression and slight overthrusting.

MECHANICS

The history of the deformation (Fig. 8) is conceived to be as follows:

The rocks of the Cumberland block were subjected to strong lateral compression applied from the southeast. The thicker sedimentary rocks west of *A* (Fig. 2, p. 355) seem not to have yielded as did those to the east, and acted as a buttress against which the rocks to the east were deformed. The compressional stress was much more intense at the Tennessee end of the block and the first result of the stress was the folding of the Powell Valley anticline and of the lateral anticline which later broke and formed the Jacksboro cross fault. Between *A* and *B* Keith[1] found evidence of this now broken anticline which was the result of deformation of the rocks of the Cumberland block against the more competent buttress on the west.

After the Powell Valley anticline had been in large measure formed and the Jacksboro cross anticline had probably reached its full development, the stresses were then transmitted across the block, and yielding farther northwest resulted in the folding of the rocks into the Pine Mountain anticline. It is probable that by this time overthrusting and shearing to the northwest had commenced at the southern end of the Jacksboro cross anticline, for the movement of the rocks of the Powell Valley anticline northwestward differentially with respect to the nearly undisturbed rocks on the west had already been very considerable. With the continued crumpling of the Pine Mountain anticline, the Jacksboro cross fault developed progressively toward the northwest, and, when it reached the then position of the corner *C* of the block, initiated the great Pine Mountain fault.

There had by this time been considerable skewing of the entire block which was pivoted at or near its north corner, with the result that the corner of the block at *A* had been thrust more extensively on the rocks to the west than had the corner at *C*. The overthrust to the west in the Jacksboro cross fault is, however, believed to be only the smaller movement incidental to the skewing of the block, while the main movement in this fault was

[1] Arthur Keith, *U.S. Geol. Survey Geol. Atlas, Briceville Folio No. 33* (1896).

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the shearing by which the block to the east moved several miles to the northwest. The writer does not believe that there was in the original stress any distinct southwesterly component, but

FIG. 6.—Slickensided shale near mouth of Lick Creek south of Birchleaf

considers the Jacksboro thrust to be solely the result of the twisting of the block.

The Pine Mountain fault is compound at *C*, consisting of four distinct overthrusts, but becomes more simple northeastward. The faulting which commenced at the southwest developed progressively toward *D* as the stress continued, but naturally the

total displacement was less at *D* than at *C*. It seems likely that there was some slight displacement beyond Skegg Gap and into Dickinson County along the Pine Mountain fault before it was intersected by the Russell Fork cross fault.

In following chronologically the development of the Jacksboro–Pine Mountain line of faulting the Russell Fork cross fault was temporarily omitted. The history of its development is correlated with the events described above, as follows:

Only after there had been considerable development of the Pine Mountain anticline at *C*, and some shearing along the Jacksboro cross fault, was the skewing of the Cumberland block felt at the northeast end. Its first expression was the development of a tension or normal fault starting at *F* (Fig. 2) and extending toward *E* with continued twisting of the block.

The presence of the Little Pawpaw fault, which appears to be primarily the result of such tension incident to twisting, leads the writer to believe that the point, or, perhaps more correctly, the area of pivoting, was somewhat to the north of *E*. On the other hand, evidence of somewhat more pronounced compression along the fault from *D*, part of the distance toward *E*, seems to indicate that compression was even at first dominant in that part of the line. It seems therefore probable that the region of pivoting is located between *D* and *E* but somewhat nearer the latter.

After the extension of the Pine Mountain fault beyond Skegg Gap and the extension of the Russell Fork cross fault beyond *E* as a normal fault, the accumulation at the northeast end of the block of the northwestward-trending stresses, which had long been operative at the Tennessee end of the block, reached the critical point, and the northeast end was broken loose along the line largely determined by the pre-existing normal fault. The line of this break intersected the Pine Mountain fault at Skegg Gap, stopped farther movement in that fault east of that line, and permitted the Cumberland block to be thrust not over two miles northwestward at this end. Since the Russell Fork fault line forms an angle of over 90° with the line of the Pine Mountain overthrust, the overthrusting of the east end of the block brought about compression along the whole extent of the Russell Fork

fault, reversing the condition of tension which produced the normal fault, and producing overthrusting and considerable crumpling of the shale and crushing and jointing of the sandstone adjacent to the fault plane. The net amount of overthrust is very slight, probably at no point reaching 500 feet, and the rocks on the southwest side of the fault are nowhere over 250 feet above those on the northeast. The shearing loose of the northeast end of the block, its overthrust along the already established Pine Mountain

FIG. 7.—View of Skegg Gap looking north along fault line. At this point the resistant basal conglomerate of the Lee on the overridden side (right) is adjacent to the weak Pennington rocks of the overthrust side (left).

fault, and the compression and slight thrusting along the Russell Fork cross fault were the closing events in the history of the Cumberland block as a unit.

There are many especial features which are particularly in accord with this interpretation of the movement of the Cumberland block as a progressive skew with final release of the east end. The skewing of the southwest end of the block first with the Skegg Gap corner remaining longest in place explains admirably the otherwise anomalous facts of rather strong overthrust in the Jacksboro fault, the trace of which is nearly at right angles to

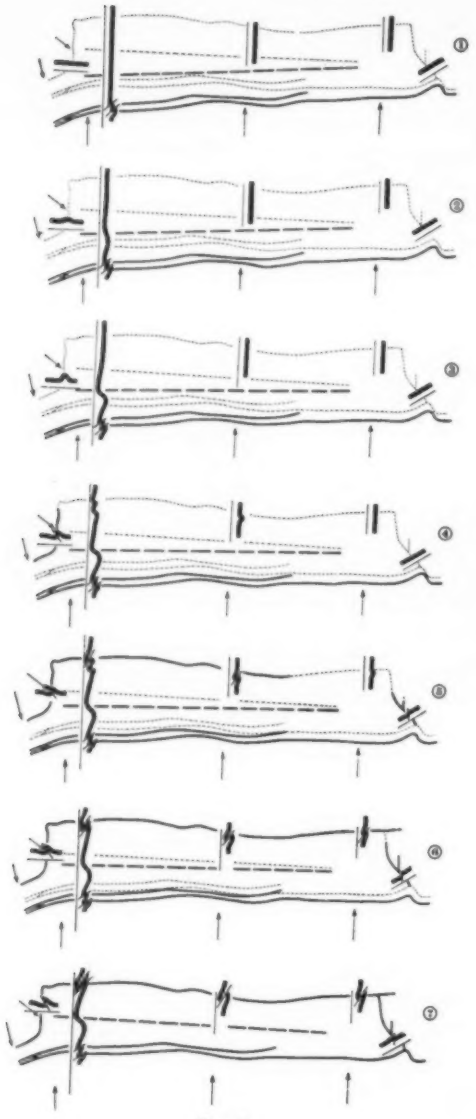

FIG. 8

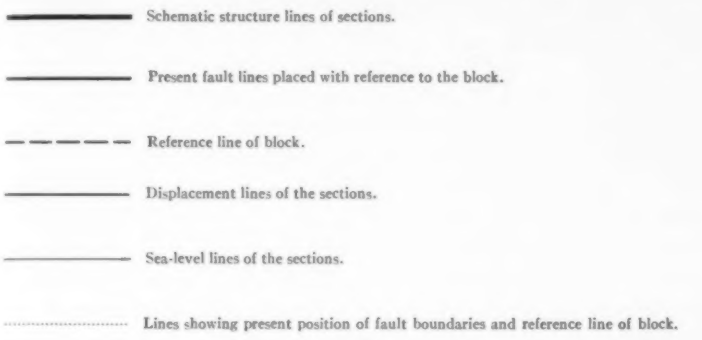
Schematic structure lines of sections.
Present fault lines placed with reference to the block.
Reference line of block.
Displacement lines of the sections.
Sea-level lines of the sections.
Lines showing present position of fault boundaries and reference line of block.

FIG. 8.—Serial diagram showing history of the deformation and displacement of the Cumberland block as interpreted by the writer. In sketch No. 7 of the series is shown the present condition in outline. The heaviest lines show the structure in the sections schematically. The lightest lines are the "sea-level" lines of the sections and the fault lines of the sections still referring to sketch 7 of the series above. The solid lines of medium weight are the present traces of the various faults. The medium-weight dotted line is an arbitrary axis line the position of which is the same with reference to the block in each sketch of the series.

In sketch 1 of the foregoing series the present fault traces are shown by the light dotted lines of the background and the present position of the axis line by the light and straight dotted line. The then position of the fault traces and of the axis line of the block is shown in sketch 1 by the medium-weight solid and dotted lines respectively. The light "sea-level" lines and fault lines of the sections and the very heavy lines of the schematic sections are the same as in sketch 7. The arrows which point upward in the sketches indicate the direction of the main thrust; the arrows pointing down and to the right at the left-hand end of each sketch indicate the resisting stress of the buttress southwest of Jacksboro as described in the text.

The first sketch assumes the prior formation of the Hunter Valley. The second shows some slight buckling of the southwest end of the block. The third shows more intense buckling here. The fourth shows more intense buckling and extension of the folding eastward along the line of the Pine Mountain fault. In No. 4 also is shown the beginning of the Jacksboro overthrust fault. In No. 5 this fault has extended far around the north side of the block and the Russell Fork fault has been initiated as a normal fault. Sketch 6 shows further extension of the faulting and only a small corner of the block near the northeast corner remains attached. In sketch 7 this small attachment is broken and the block has been thrust into its present position. The relative movement of the block at different stages has been shown by the gradual migration of the heavy dashed axial line toward its final position as shown by the light dotted axial line of each sketch.

the trace of the Pine Mountain fault and of only slight overthrust and more restricted compression in the Russell Fork fault with its trace at a much greater angle with the main overthrust.

The stronger development of the Powell Valley anticline and the presence of the Wallen Valley fault only at the southwest are also strongly in accord with this suggested interpretation.

The greater intensity of stress implied by the compound character of the Pine Mountain overthrust at *C* (Fig. 2) and the probable development of the Powell Valley anticline before the rocks of the block were competent to transmit the stresses to the Pine Mountain fold which later broke in a fault, point strongly to the initiation of the faulting at the south end of the Jacksboro cross fault to allow the necessary shortening of the strata on the northeast side of that fault. The four faults as interpreted by Keith, and corroborated by the displacement of the north limb of Powell Valley anticline in the Briceville quadrangle, give clear evidence of a movement of at least ten miles to the northwest. At Skegg Gap the evidence does not indicate over two miles of overthrust at the most.

In the course of his meditation on this study the writer has made very briefly a few computations, based on extremely general and only very approximate assumptions, which are given below. Their value is solely to indicate orders of magnitude, and it is hoped that they may serve, as they did in the case of the writer, to visualize the immensity of forces involved.

FORCE TO SHEAR AND FORCE TO THRUST

Assumptions

Block 125 miles $\times$ 25 miles $\times \frac{1}{2}$ mile
Density 170 lbs. per cubic foot
Coefficient of friction, mean between rough and smooth granite, 0.60
Shearing strength 200 pounds per square inch
Average extent of overthrust, 6 miles

Results

Force to shear block loose over entire area $= 25 \times 10^{14}$ pounds
Force to move block against friction on horizontal plane $= 23 \times 10^{15}$ pounds
Work done in moving block 6 miles at angle of 5 degrees $= 85 \times 10^{19}$ foot pounds
Equivalent to 420,000 horse-power working for 100,000 years
Estimated coal in block $= 50 \times 10^{9}$ tons

Burning of this coal would produce power enough to move the block 2.2 feet, assuming the usual engine efficiency. It has actually moved an average of at least six miles.

It is especially interesting to note that the force required to shear the block loose over the whole area is only about one-tenth of that required to produce motion against the resistance of friction. Since both forces are proportional to area and only one—that of motion against friction—proportional to thickness, we find that for a block of any area and of a thickness of 287 feet, according to the conditions assumed, the shearing force is just equaled by the force to overcome friction, and as thickness is greater than this amount the latter force is greater in proportion. It is evident, then, that in the case of most overthrust faults the motion of the rock involved against the resistance of friction is more impressive than the production of the break which separated it.

STUDIES OF THE CYCLE OF GLACIATION

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University of Michigan

I. THE CYCLE OF MOUNTAIN GLACIATION WITHIN MODERATE LATITUDES

In a general discussion of glacial sculpture in mountains,[2] the writer has made use of the terms *grooved* or *channeled upland* and *fretted upland* to describe respectively the early and the late effects of the erosional action of mountain glaciers. The Bighorn Mountains in Wyoming and the Swiss Alps were chosen as type examples of these contrasted erosion surfaces, the characteristics of which are, that in the former large areas of the preglacial upland still remain (Figs. 1, 2, and 3), while in the latter its complete dissection by cirque recession and enlargement has resulted in a system of main and secondary rock palisades described as comb ridges. Between these contrasted land surfaces many gradations exist, though examples of the former are relatively uncommon. Similar to the channeled upland of the Bighorn Mountains, though with less of the preglacial surface retained, are portions of the Uinta and Wasatch mountains, of which excellent illustrations have been supplied by Atwood.[3] All the best examples are furnished by the Rocky Mountains in the interior of the American continent, where the moist westerly winds have been robbed of their moisture in crossing the high Sierra Nevada and Cascade ranges.

In a visit to the Glacier National Park, the writer was impressed with the fact that a type of topography is there represented which indicates a still later stage of sculpture by mountain glaciers than

[1] Illustrations from photographs by the author.

[2] "The Cycle of Mountain Glaciation," *Geogr. Jour.*, Vol. XXXV (1910), pp. 147–53, Figs. 1–19. Also, "Characteristics of Existing Glaciers," pp. 25–40, Pls. 3–9.

[3] "Glaciation of the Uinta and Wasatch Mountains," *U.S. Geol. Survey, Prof. Paper 61* (1909), maps of Pl. 8A.

does the fretted upland as exemplified by the Alps. The most striking peculiarities of this type are found in the unusual number of isolated sharp peaks of monumented aspect (Figs. 4 and 5), and this is combined with a general absence of the comb ridge (a rare example is shown in Fig. 6) and a frequency of unusually low cols or passes (Fig. 7). Unlike the true horns of the Matterhorn type, which in the fretted upland are relatively few in number and may perhaps represent by their summits points near the original surface of the upland, the monuments of the northern Rocky Mountains show a tendency to appear in pairs, and in many

FIG. 1.—View of Mt. Mathews, Bighorn Range, taken from the southeast and showing the character of the preglacial surface. At the left in middle distance is a cirque.

instances at least they are remnants of lower portions of the preglacial surface (Figs. 5, 8).

Both in the Bighorn Range and in the Glacier National Park the glaciers have today nearly or quite disappeared, being now represented by small horseshoe or cliff glacierets only. The earlier conditions of nourishment were, however, as we know from more or less extended studies, notably different from those of today. In the Bighorn Range the glaciers of Pleistocene time extended far down the valleys, where strong terminal moraines are found to mark the limits of their advance.[1]

[1] R. G. Salisbury, "Cloud Peak–Sheridan Folio," *U.S. Geol. Survey;* also, N. H. Darton, *U.S. Geol. Survey, Prof. Paper 51*, pp. 71–91, Pls. 37–36.

FIG. 2.—View of cirque above Seven Brothers Lakes, seen from the ridge south of Trail Lodge, Bighorn Range.

FIG. 3.—Nearer view of the cirque shown in Figure 2. Characteristic surface of preglacial area in foreground.

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In the Glacier Park district the Pleistocene glaciers occupied the entire valleys within the range and spread out eastward their aprons of Piedmont type. They also extended westward a long distance down the valley of the Flathead River.[1]

It is here proposed to use the term *monumented* upland to describe the extreme type of mountain sculpture which is represented in the Glacier National Park and which is believed to be due to continued glacial action upon a fretted upland like that of the Alps. Cirque enlargement carried to this stage has sapped the

FIG. 4.—View of Reynolds Mountain, a characteristic monument of the Glacier National Park region. View taken from the trail to Piegan Pass.

main comb ridge so as to largely obliterate the *aiguille* type of crest or *arête* (Fig. 6 shows one of the remaining comb ridges). Matterhorns have in the process been reduced in size as the cols are progressively lowered and widened and are transformed into *arêtes*. The last remnants of the upland to be removed by this continued cirque enlargement are found away from the original divide and outward toward the flanks of the upland, for the reason that in their later stages cirques enlarge excessively on their lateral walls. A good illustration of this tendency is supplied by the

[1] Wm. C. Alden, "Pre-Wisconsin Glacial Drift in the Region of Glacier National Park," *Bull. Geol. Soc. Am.*, Vol. XXIII (1912), Pl. 37. See also by same author, "Glaciers of Glacier National Park," and especially the map opposite p. 32.

FIG. 5.—A pair of monuments of monumented upland seen from the Piegan Trail, Glacier National Park.

FIG. 6.—View of a comb ridge looking across the Piegan Pass, Glacier National Park.

gently sloping summit plane of Quadrant Mountain in the Yellowstone National Park at an elevation of between 9,000 and 10,000 feet[1] (Fig. 9), since Antler Peak and Bannock Peak guard the entrance to the cirque.

It is especially because the comb ridges in the highest levels are precipitous and correspondingly thin that a continuation of

FIG. 7.—Gunsight Pass, seen from Gunsight Chalets looking across Gunsight Lake, Glacier National Park.

the process removes their pinnacles while the broader ridges somewhat farther out and just below the mother-cirques are being sharpened into peaks, both alike through sapping from the cirques.

To bring together the extremes of mountain glacier erosion which are represented by the Bighorn Range and the Glacier National Park with the intermediate stages which connect them, the four generalized plans of Figure 10 have been prepared. In order, these are:

I. The youthful channeled or grooved upland
II. The adolescent early fretted upland
III. The fretted upland of full maturity
IV. The monumented upland of old age

[1] "The Cycle of Mountain Glaciation," *Geogr. Jour.*, Vol. XXXVI (1910), Figs. 8, 14.

These four stages are perhaps best illustrated by the Bighorn Range, the mass of Snowden in the Welsh highland, the Alps, and Glacier National Park (Fig. 10).

FIG. 8.—Monuments on either side of entrance to cirque above Ptarmigan Lake, Glacier National Park. (Photograph purchased of Northern Pacific Railway.)

The two districts which are here contrasted, the Bighorn Range and the Glacier National Park, furnish also the opportunity to contrast the effects of rock structure in modifying the forms of relief shaped by mountain glaciation. Whereas the high upland of the Bighorn Range has a core of massive rock, thus resembling the Wasatch and Uinta ranges and the Alps, the rocks of Glacier National Park are sediments and dominantly shales and limestones. It was to be expected that the characteristic structures of these

FIG. 9.—Cirque at head of Panther Creek, Yellowstone National Park, with pair of monumented peaks at entrance—Antler Peak and Bannock Peak.

sediments, their bedding planes and their joint system, should exert a strong influence upon the topographic forms produced, as indeed they have. The influence of the bedding planes is displayed in the Glacier National Park in the accentuation of the rock terraces at the upper ends of valleys within the cirques themselves. As in the Canadian Rockies across the international boundary, this character reaches an extreme (Fig. 11). An excellent instance is shown also in an illustration by Alden.[1]

[1] *Glaciers of Glacier National Park*, Fig. 11.

To the well-developed jointing found in the rocks of the Glacier National Park must be ascribed the well-marked checkerboard pattern displayed by the park valleys, a pattern which strikes one at once when the topographic map is examined. A number of observations of the bearings of master-joints which were made by

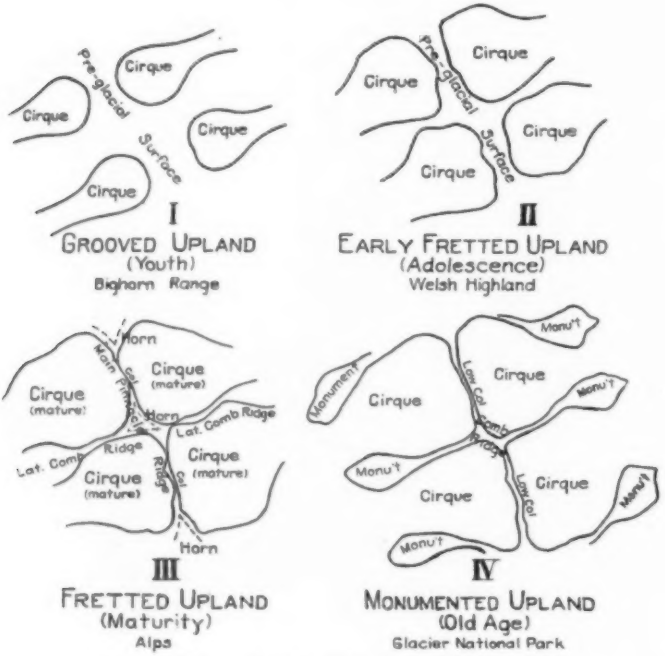


FIG. 10.—Stages of sculpture by mountain glaciers

the writer indicated a rather general correspondence between them and the trends of the valleys in which they were found. In some instances the lower spurs which have been less extensively sapped by glacial erosion indicate very clearly the dominating influence of the joint planes in shaping them. The cirques themselves also display this tendency by their approach to rectangular outlines.

II. THE TRANSITIONS BETWEEN THE MOUNTAIN GLACIER AND THE ICE CAP

From the standpoint of the sculpturing of the lithosphere, the ice cap is sharply set off from all types of mountain glacier through its inability to accomplish a sapping of rock surfaces due to rapid frost-weathering. Its sculpturing processes are therefore restricted to plucking, abrasion, and to a very limited extent frost-weathering on flattish surfaces—processes which in combination leave the rock rounded and presenting surfaces which are flatly convex

FIG. 11.—View of terraced cirque above Lake Grinnell, Glacier National Park

skyward. That these processes combined play but a subordinate rôle to frost-weathering in the case of all the types of mountain glaciers, would seem to be sufficiently attested by the sharply accented features which are brought about with their concavities toward the sky.[1]

Since the mountain glacier owes its very existence to a rock container within the lithosphere surface, the inclosing rock walls

[1] Hobbs, *Earth Features*, etc. (1911), p. 379, Fig. 405.

must in general project above the ice of the glacier. The rock surface will also be reached by air and water wherever crevasses descend through the ice of the glacier to the bed upon which it rests. The conditions essential to the sapping process are a supply of water on the rock surface and oscillations of temperature about the freezing-point. These conditions are not realized either in the case of ice caps or of continental glaciers, save only where nunataks emerge from beneath the ice near to the glacial margin.

When during an advancing hemicycle of glaciation a mountain glacier is so amply nourished that the rock walls of its containing basins become entirely submerged (ice-cap stage), a profound and immediate transformation takes place in the sculpturing processes

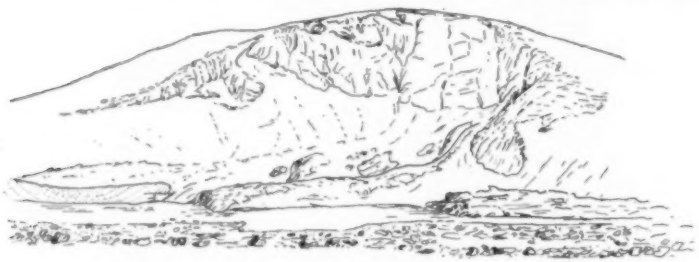


FIG. 12.—The northern cirque (Kjedel) on Galdhöpig in the glaciated surface of Norway. (After E. Richter.)

Up to this time, under the dominating influence of the sapping process, the effect of the glacial sculpture has been to sharpen all projecting features of the relief as the glacial basins and channels are carved deeper and extended outward from each individual locus. Now, however, under the plucking and grinding processes alone, which have usurped the functions of the frost-weathering, the pinnacles and horns within the comb ridges are truncated and ground down, with the result that above the shallowed cirques and the largely obliterated U-valleys there extends a flatly convex surface like that which is fashioned by the same processes beneath a continental glacier. The sharp relief which was inherited from the period of mountain glaciation is thus gradually ironed out into a flatly convex surface which is everywhere ground and polished by abrasion. The U-valleys are first effaced, beginning at their

lower extremities, and the last of the hollowed features of the inherited surface to disappear are the increasingly truncated remnants of the cirques, which in their later stages take the form of an armchair-like depression. Such features are well displayed in Norway where the continental glacier has similarly ironed out the inherited grooved or fretted upland (Fig. 12).

Such a surface as succeeds to a fretted upland under the sculpturing action of either an ice cap or a continental glacier will resemble in form a grooved glacial upland of extreme youth such as is illustrated by the Bighorn or Uinta ranges of the Rocky Mountain region, but it has less pronounced relief and, unlike such a preglacial remnant ("biscuit-cut" surface), *the upwardly convex surfaces are here planed and polished by abrasion.*

In the receding hemicycle of glaciation which succeeds to the culmination of glacial alimentation, the flat dome of the ice which constitutes the ice cap will have its surface progressively lowered until the stage is reached at which the rims of the buried cirque remnants begin to emerge from beneath their mantle of ice. In West Antarctica, near the winter quarters of the Swedish Antarctic Expedition of 1901–3, ice caps now blanket both James Ross and Snow Hill islands, and, like all Antarctic glaciers, they are in a receding hemicycle of glaciation. On the first-named island the rims of the cirques have emerged from beneath their cover along the eastern and southern margins of the island (Fig. 13). The Gourdon and Rabot glaciers are already apparently in large part detached from the dome of the ice cap, which here rises to its highest point in the Haddington berg. In the largest of the cirques lies the Hobbs Glacier, which is still in part fed by two ice cascades situated near the middle of the rim.[1]

Except that the continental glacier, and not an ice cap, has been the modeling agent, Mount Washington in the White Mountains[2]

[1] Otto Nordenskjöld, "Die schwedische Südpolar-Expedition und ihre geographische Tätigkeit," *Schwedische Südpolar-Expedition, 1901–3*, Vol. I, Lieferung 1 (1911), pp. 154–55, Map 3 and Pl. 13, Fig. 1.

[2] J. W. Goldthwait, "Following the Trail of the Ice Sheet and Valley Glacier on the Presidential Range," *Appalachia*, Vol. XIII (1912), pp. 1–23 (reprint), Pls. 1–9; "Glacial Cirques Near Mt. Washington," *Am. Jour. Sci.*, Vol. XXXV (1913), pp. 1–19; "Remnants of the Old Graded Upland on the Presidential Range of the White Mountains," *ibid.*, Vol. XXXVII (1914), pp. 451–53; "Glaciation in the White Mountains of New Hampshire," *Bull. Geol. Soc. Am.*, Vol. XXVII (1916), pp. 263–94, Pl. 13.

and Mount Ktaadn in Maine[1] would appear to supply near parallels to the sculpture just described, since the "gulfs" of the districts have been clearly recognized as cirques. Tarr has claimed that the mountain glaciers which sculptured the cirques on Mount Ktaadn

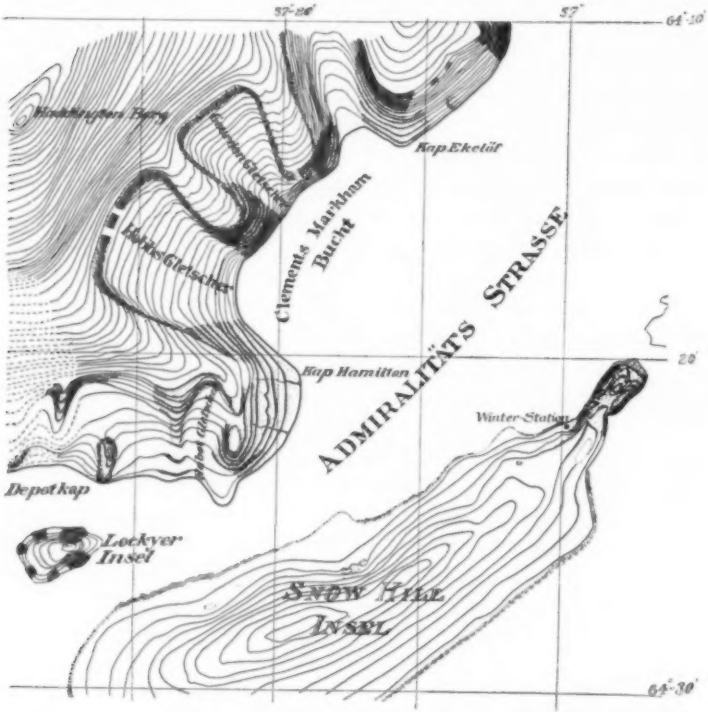


FIG. 13.—Map of portions of the James Ross and Snow Hill Islands of West Antarctica. (After Otto Nordenskjöld.)

were subsequent to the continental glaciation of the region. This is disputed by Goldthwait, who brings forward evidence to prove that in the White Mountains the mountain glaciers were antecedent to the continental glaciation which shaped the higher and flatter rock surfaces. We hardly see how there could fail to be glacial

[1] R. S. Tarr, "The Glaciation of Mt. Ktaadn," *Bull. Geol. Soc. Am.*, Vol. XI (1900), pp. 433–48.

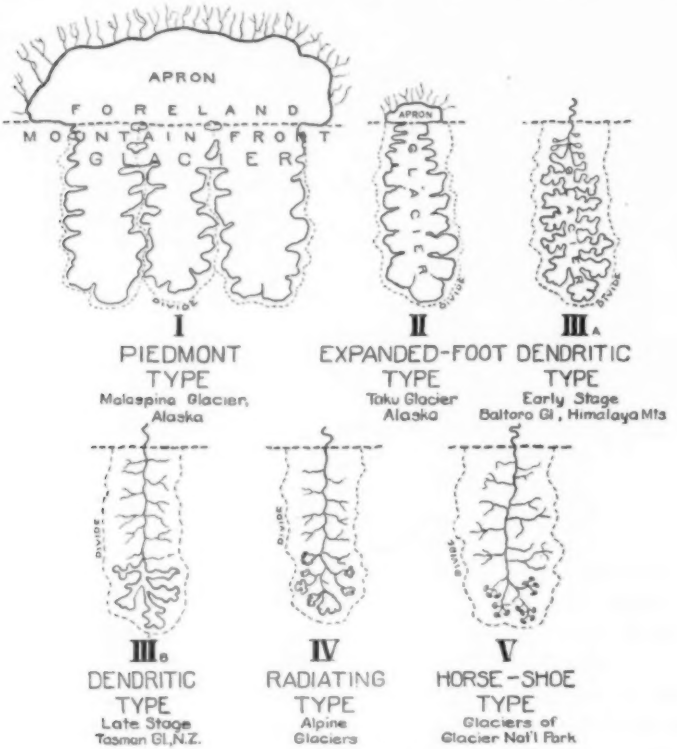
remnants in occupation of the cirques for at least a brief period while the continental glacier was withdrawing from the region. These would presumably develop in much the same manner as those already described on James Ross Island, but with differences which will be pointed out in the next section of this paper. Goldthwait is no doubt correct in believing that the mountain glaciers had a much longer life during the advancing hemicycle of glaciation and that the cirques were shaped at that time. It is even doubtful if any appreciable work of erosion or deposition was accomplished in the later period of mountain glaciation, and this interpretation would be in harmony with Goldthwait's observations.

III. THE GLACIAL CYCLE ON THE MARGINS OF THE CONTINENTAL GLACIER OF ANTARCTICA

It is a fundamental and prerequisite condition for the sequence of stages through which mountain glaciers pass during a receding hemicycle of glaciation that the areas of alimentation and ablation should be sharply separated from each other. The former is restricted to the upper levels, and alimentation is augmented in amount toward the top, whereas the area of wastage is found in the lower levels and the losses are increased toward the bottom. Such a distribution results principally from two conditions: (1) mountain glaciers are nourished by upwardly directed air currents which deposit their moisture as a result of progressive adiabatic refrigeration; and (2) they are wasted by contact with warm-air layers whose temperature rises progressively toward the bottom. *It is a direct consequence of the combination of these conditions that mountain glaciers during a receding hemicycle of glaciation become reduced in area through withdrawal of the glacier foot up the valley, and even in its expiring stage the glacier head occupies essentially the same position that it did at the beginning* (Fig. 14).

Were these two conditions affecting the size of mountain glaciers not realized, the results would be quite different. When we examine the glaciers on the margins of the inland ice of the Antarctic, we find they differ widely from those of moderate latitudes, which are the ones that are well known and have formed the basis of our classification. Within the Antarctic air temperatures do not rise

above the freezing-point even in the summer season, save only during short intervals at the termination of the fierce Antarctic blizzards. Furthermore, these marginal glaciers to the inland ice are nourished, not by inwardly and upwardly directed air currents,



GLACIER TYPES OF RECEDING HEMICYCLE
Progressive Withdrawal of Glacier Foot

FIG. 14

as are the mountain glaciers of moderate latitudes, but by downwardly and outwardly flowing currents which bring drift snow from the inland ice and often carry it beyond the marginal glaciers to be dissipated upon the surface of the sea. *Separate areas of*

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FIG. 15.—Map of an area in South Victoria Land near the winter quarters of the last Scott Expedition, showing waning glaciers which are withdrawing at both their upper and their lower margins. (After Griffith-Taylor and others.)

nourishment and waste in distribution with reference to altitude are thus not realized, and the otherwise universal law of exclusive drawing in of the foot of the glacier during its waning stages does not hold.

That this is true is particularly well shown in the area of waning glaciers described as "ice-slabs" by Ferrar, the glacialist of the first Scott Expedition to the Antarctic, and fully mapped by Griffith-Taylor, Debenham, and Wright of the last Scott Expedition[1] (Fig. 15). On a far larger scale and related to a continental glacier rather than an ice cap, these dying glaciers represent a later stage than those marginal types which have already been referred to from West Antarctica—the Gourdon, Hobbs, and Rabot glaciers of James Ross Island.

By examination of the map (Fig. 15) it will be noted that these glaciers must in an earlier stage have been connected together as a piedmont which was then a part of the parent area of inland ice lying to the westward. From that continental glacier when detachment occurred the rims of the battery of remodeled cirques which rise west of the existing glaciers must have emerged from the ice mantle in forms not unlike those now seen on the margins of James Ross Island. *Their subsequent diminution in size has gone on through withdrawal both from the cirques and from the lower portions of their valleys—from both extremities toward a central position at a moderate altitude, where the last stand will be made before final extinction.*

The usual law of ablation regulated with respect to altitude here plays, therefore, no part, and it is evident that the reflection and consequent intensification of solar heat radiation in the neighborhood of exposed rock walls has here been the controlling factor in localizing the wasting process. This effect of exposed rock surfaces has been recognized for high latitudes by the observation of moats surrounding nunataks[2] and of the lateral streams beside glacier tongues[3]

[1] Robert F. Scott, *Scott's Last Expedition*, Vol. II, map opposite p. 198.

[2] "Characteristics of Existing Glaciers," pp. 169, 257, Pl. 33B.

[3] *Ibid.*, Pl. 25A.

Reviews

Two Gas Collections from Mauna Loa. By E. S. Shepherd. *Bull. Hawaiian Observatory*, Vol. XIII, No. 5, May, 1920.

This is a brief report by Dr. Shepherd of the Carnegie Geophysical Laboratory at Washington on two gas samples collected by Dr. T. A. Jaggar, Jr. The samples were taken near the edge of a flow of incandescent rough pohoehoe lava on the south slope of Mauna Loa. The gases were collected in vacuum tubes from a depth of 2 feet in a 2-inch crack in the lava surface. The lava at a depth of 3 feet was glowing and the estimated temperature at the point of collection was 300° C.

A condensation of water within the tube was noted immediately upon collection. The analyses showed that about 70 per cent by volume of the gas (computed at 1200° C. and 760 mm. pressure) was water, in which respect the gases closely resemble those of Kilauea. About 16 per cent was nitrogen and the remainder mainly SO_3, SO_2 and CO_2. The water cannot be explained as the result of oxidation of hydrogen by admixed air, as is shown by the nitrogen percentage. If all the nitrogen were assumed to come from admixed air, the oxygen in such a quantity of air would be insufficient to account for the observed water.

The evidence of these samples accords with the classic work of Day and Shepherd at Kilauea in demonstrating the abundant presence of water in certain volcanic gases.

The gases of Mauna Loa show a high degree of oxidation, i.e., they have been almost completely burned. In general, they show a high degree of similarity to the Kilauea gases although the latter are rather variable. Especially noteworthy at Mauna Loa is the abundance of SO_3—2 to 8 per cent.

E. S. Bastin

The Geology and Ore Deposits of the Virgilina District of Virginia and North Carolina. By Francis Baker Laney. Virginia Geological Survey, University of Virginia. (Prepared jointly by the Virginia Geological Survey and the North Carolina Geological and Economic Survey.) 1917. Pp. 176.

The Virgilina district which lies partly in Virginia and partly in North Carolina is one of the copper districts in the eastern United States

that has produced considerable tonnages of ore. The investigation covers an area of approximately 550 square miles, including parts of Charlotte, Halifax, and Mecklenburg counties, Virginia, and parts of Granville and Person counties, North Carolina.

The area is made up almost wholly of igneous and highly metamorphosed rocks. They include ancient metamorphic gneisses and schists, the origin of which is unknown; a sequence of volcanic rocks, both basic and acidic, and volcanic clastics of each type, together with much volcano-sedimentary material; intrusive rocks of both basic and acid types, such as gabbro, diorite, granite, and syenite; and different varieties of dike rock, especially diabase. There is a small area of red or brown sandstone of Triassic-Newark age. Except the intrusives, the sandstones, and the dikes the rocks are all highly schistose and gneissoid in texture.

This prominent schistosity of the rocks is probably the most obvious structural phenomenon of the district, although jointing is prominent and there is conclusive evidence of folding. There is little direct evidence of faulting, but the intense dynamic metamorphism of the district could hardly have occurred without causing a certain degree of faulting.

With the exception of a few mineralized areas in more or less epidotized zones of the true basic schist, where deposits of native copper or of cuprite occur, the ore deposits are found in well-defined fissure veins, which occupy fractures in the rocks—in some instances possibly fault planes. The rock in which the veins occur is basic in character—the Virgilina greenstone—having the mineralogical and chemical nature of andesite; but it is thought that the vein material, both ore and gangue, was derived from the granitic magma of the region.

The gangue minerals, exclusive of included fragments of schist, named in the approximate order of their abundance, are: quartz, calcite, epidote, chlorite, hematite, sericite, albite, and possibly other plagioclase feldspars in small amount, and pink orthoclase.

The ore minerals, named in the approximate order of their abundance are: bornite, chalcocite, native copper, malachite, azurite, cuprite, chalcopyrite, chrysocolla, klaprothite (?), pyrite, argentite, silver, and gold. Of these minerals, bornite (in part), chalcocite (in part), chalcopyrite (in part), pyrite, klaprothite, argentite, native copper, and gold are regarded as hypogene or primary; while a part of the chalcocite, bornite, and chalcopyrite, and all the native silver, cuprite, malachite, azurite, and chrysocolla are held to be supergene or secondary.

The author gives a description of individual mines and prospects. A good geologic map of the district is appended to the report.

R. A. J.

The Stratigraphy and Correlation of the Devonian of Western Tennessee. By CARL O. DUNBAR. State of Tennessee, State Geological Survey, Bull. No. 21, Nashville, Tenn., 1919.

This volume is a detailed statement of the stratigraphy and correlation of the Devonian rocks of the western valley of the Tennessee River. The long sequence of the Devonian strata exposed in this region, especially the presence of the Upper Oriskany, and the abundance of fossils, probably will make this the standard section of the Lower Devonian of the entire Mississippi Basin. The important paleontological aspects of the problem are well treated. Following is the sequence of the Devonian formations of western Tennessee, as given by the author:

Series	Group	Formation
Neo-devonian	Chautauquan	Chattanooga shale Hardin sandstone member
	Senecan	Break
	Erian	
Meso-devonian	Ulsterian	Pegram limestone Break Camden chert Break
	Oriskanian	Harriman chert Break Quall limestone
		Break
Paleo-devonian	Helderbergian or Linden	Decaturville chert Break Birdsong shale Break Olive Hill formation: Flat gap limestone; Bear Branch limestone; Pyburn; Ross limestone Break Rockhouse shale

R. A. J.

The Geology and Coal Resources of the Coal-bearing Portion of Tazewell County, Virginia. By T. K. HARNSBERGER. Virginia Geological Survey, University of Virginia, Bull. No. 19. Prepared in co-operation with the U.S. Geological Survey. 1919. Pp. 195.

This report deals with the coal resources of Tazewell County in southwestern Virginia. The surface rocks in the coal district belong to the Devonian, Mississippian, and Pennsylvanian systems. All the commercially valuable coal is in the Pennsylvanian.

The most prominent structural feature of the area is the Dry Forks anticline. The Pocahontas syncline and other folds occur in the region. The coal area is bounded on the southeast by a series of thrust faults. The Tazewell County coal field originally extended to the southeast far beyond its present limits but folds and faults lifted the coal-bearing rocks of the region to the southeast far above those of the present field and they have been removed by erosion.

The total area of coal land is 696.5 square miles. The total thickness of the coal-bearing formations is about 2,800 feet, every portion of which is exposed in some part of the area. At least fifteen coal beds are 30 inches or more in thickness over territory of sufficient extent to justify mining. In general the coal is of good coking quality and has a high fuel value. Because of the extreme variability of the coal beds, plans for development should be preceded by careful geological examination. Complete descriptions of the various coal beds are given. Included in the report are both a topographic and a geologic map of the coal area.

R. A. J.